# Geology of Ontario

## Ontario Geological Survey

Special Volume 4, Part 2

Ministry of
Northern Development
and Mines
Ontario

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## The Cover:

The cover shows the Niagara Escarpment overlooking Georgian Bay at the Lion's Head Provincial Nature Reserve on the Bruce Peninsula. The escarpment, an erosional cuesta extending from western New York State through Ontario to northern Michigan, is the most significant physiographic feature in southern Ontario and the best exposure of Paleozoic rocks in the province. The Bruce Trail runs along the Niagara Escarpment in Ontario, providing access to well-exposed geological features, unique natural environments and many marvellous vistas.

At the photo location, clastics of the Lower Silurian Cabot Head Formation form the narrow ledge just below the water level and are obscured by talus on the vegetated slope beneath the cliff. Interbedded shales and carbonates of the Dyer Bay, Wingfield and Fossil Hill formations underlie the vegetated part of the slope. Overlying these rocks, the white cliffs of the escarpment consist of thick- to massive-bedded dolostones of the Silurian Amabel Formation. This formation consists mainly of sand-sized carbonate grains derived from the mechanical breakdown of shells and skeletons of crinoids, corals and brachiopods which thrived in the warm tropical sea covering this area approximately 425 million years ago. Dimension stone quarried from the Amabel Formation has been used for construction of the Canadian Chancellery in Washington, D.C. and the Provincial Courthouse in Ottawa.

The main direction of glacier flow in this area during the Quaternary was to the [illegible]-southwest, roughly parallel to the cliffs in this photo. Scattered evidence of ornate sculpting of the cliff face by tur[illegible]ubglacial meltwaters is preserved. Postglacial weathering has removed or buried much of the evidence of the passage of continental glaciers over the Bruce Peninsula. The once glacially polished and finely striated surface of the carbonate rock has beeen naturally corroded through dissolution of carbonate minerals by surface and groundwaters. These chemical weathering processes form a karst landscape with dissolution-enhanced joints and fractures, shallow grooves on the bedrock surface (bottom right of photo), and sinkholes. Excellent examples of these karst forms occur on Lion's Head promontory and at many other localities on the Bruce Peninsula.

Cover Photo: P.S.G. Kor, summer 1991
Caption: D.K. Armstrong, P.J. Barnett and P.S.G. Kor
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Ontario Geological Survey

Special Volume 4
Part 2

# Geology of Ontario

Edited by

P.C. Thurston, H.R. Williams,
R.H. Sutcliffe and G.M. Stott

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# Foreword

*Geology of Ontario*, built upon 100 years of mapping and research into the geology of the Province, represents the collected geoscientific expertise of the Ontario Geological Survey. It also embodies the skills and talents of numerous people, from authors, editors, cartographers and drafters to laboratory technicians, computer specialists, research assistants, typesetters, printers and administrative staff. These people, many of whom were brought together especially for this project, came from government and industry across Ontario to provide the craftsmanship so vital to a project of this magnitude.

*Geology of Ontario* is the most ambitious project in geoscience publishing yet undertaken in Ontario, and possibly Canada. Its 41 sheets of maps and charts, and 1430 pages of text and illustrations are the visible result of a creative process which started just over five years ago, and a production effort which was the focus of our publishing team for the last two years. The production demands of the undertaking accelerated our move into a new era of information management and dissemination, and led to the development of new technology for publishing maps. A leading example of this is the special software, developed through the co-operation of Ontario Geological Survey cartographic and scientific staff with private sector service companies, which allowed the printing of digitally coloured maps by mass-printing methods for the first time.

The long, difficult, occasionally exasperating, but always interesting job of bringing *Geology of Ontario* to life has provided an opportunity to reflect upon and consolidate our geoscientific understanding of the Province. *Geology of Ontario* will stand as a valuable reference work for many years before conceding naturally to the advance of science. Equally remarkable, these publications mark a watershed in geoscientific information distribution; starting in the conventional and migrating, through innovation and adaptation, to new computer based production and printing methods.

As a result, the OGS moves on from the first hundred years, not only with pride in past achievements, but with scientific and communication experience and skills and an open-minded attitude to innovation which should serve well in meeting the challenges of the second century.

**V.G. Milne**
***Director***
***Geoscience Branch***
***Ontario Geological Survey***

# Grenville Province

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## Chapter 19

# The Grenville Province and the Proterozoic History of Central and Southern Ontario

**R.M. Easton**

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## Abstract

The Grenville Province is a complex orogenic belt of *circa* 1.1 billion years in age that truncates several older geologic provinces. Recorded within the Grenville Province is the tectonic evolution of the southeast margin of Laurentia during the Mesoproterozoic. The Grenville Orogeny has overprinted the structural trends and metamorphic effects of the Archean and Paleoproterozoic geologic provinces of Laurentia.

The Grenville Province is subdivided, from north to south, into the Grenville Front Tectonic Zone, the Central Gneiss Belt and the Central Metasedimentary Belt. The Grenville Front Tectonic Zone consists of deformed and metamorphosed rocks derived from the Southern and Superior provinces, and the Killarney Magmatic Belt, within 20 to 30 km of the Grenville Front. Rocks of the Killarney Magmatic Belt intrude the 2.4 billion-year-old Huronian Supergroup and are 1750 to 1730 and 1470 to 1450 million years old. Southeast of the Grenville Front, part of the Killarney Magmatic Belt was affected by the Grenville Orogeny and is thus included within the Grenville Province. The Killarney Magmatic Belt is interpreted to be an extension of one of several Mesoproterozoic orogens underlying much of the North American midcontinent.

Except near Manitoulin Island, basement rocks in southwestern Ontario have all been affected by a *circa* 1100 to 1070 Ma orogenic event, the "Grenville Orogeny". The Grenville Front marks the northwesterly limit of penetrative Grenvillian metamorphism and deformation and truncates structural trends in the adjacent structural provinces, although locally some of these rocks can be traced across the Grenville Front. It is now generally accepted that this orogenic event (or events) involved northwest-directed thrusting and imbrication of the entire crust, presumably as a result of a terminal collision with a continental landmass at about 1100 Ma.

The Central Gneiss Belt consists mainly of upper amphibolite and local granulite facies, quartzofeldspathic gneisses, chiefly of igneous origin with subordinate paragneiss. Distinctive lithotectonic terranes, some further subdivided into domains, have been identified within the Central Gneiss Belt. The terranes and domains are distinguished by differences in rock types, internal structure, metamorphic grade, geologic history and geophysical signature and are bounded by zones of intensely deformed rocks traceable for tens of kilometres. Similar terranes and domains can also be identified in the subsurface of southwestern Ontario where they influenced Paleozoic sedimentation. The terranes form the basis for the description of the geology and mineral deposits of the Central Gneiss Belt.

Rocks of 3 main ages are present in the Central Gneiss Belt. North of the French River, reworked Archean and Paleoproterozoic gneisses of the Nipissing Terrane are intruded by several ages of Mesoproterozoic plutonic rocks, with granitic and monzonitic rocks predominant. The bulk of the Central Gneiss Belt (Algonquin and Tomiko terranes) consists of 1.8 to 1.6 billion-year-old gneisses intruded by 1.5 to 1.4 billion-year-old granitic and monzonitic plutons that may represent an extension of the Eastern Granite–Rhyolite Province across the Grenville Front. The Tomiko Terrane may consist entirely of Mesoproterozoic rocks that are allochthonous with respect to the adjacent Nipissing Terrane. The Algonquin Terrane probably represents a section of 1.8 to 1.6 billion-year-old crust extensively injected by granitic magmas of the Eastern Granite–Rhyolite Province. Although imbricated by later thrusting, the Algonquin Terrane is probably parautochthonous.

Overlying and southeast of these predominantly parautochthonous terranes, throughout the Grenville Province in Canada, are several allochthonous terranes. Within Ontario only 1 allochthonous terrane, the Parry Sound Terrane, has been identified. It consists of mafic to intermediate rocks with a Sm-Nd model age of about 1.45 to 1.35 Ga and in comparison to the adjacent Algonquin Terrane, it consists entirely of juvenile crust. The Fishog Domain (Algonquin Terrane), characterized by an intense aeromagnetic high and a distinctive plutonic suite, may also be allochthonous.

The Central Metasedimentary Belt is a major Mesoproterozoic accumulation of supracrustal rocks invaded by compositionally diverse, syntectonic, late tectonic and posttectonic plutonic rocks. The entire succession has been metamorphosed at grades varying from greenschist to granulite facies. The Central Metasedimentary Belt can be divided into several lithotectonic terranes. The detailed geology of these terranes and their mineral deposits are detailed in this chapter. In addition, stratigraphic nomenclature throughout the Central Metasedimentary Belt has been compiled and has been revised where necessary with the introduction of several new stratigraphic units.

Two discrete regions exist within the Central Metasedimentary Belt. A superterrane composed of the Bancroft, Elzevir, Mazinaw and Sharbot Lake terranes, including the classical Grenville Supergroup, characterized by volcanism and sedimentation between 1300 and 1250 Ma, followed by plutonism at 1250 to 1230 Ma, and metamorphism and deformation at 1230 to 1180 Ma and at 1120 to 1070 Ma. The Frontenac Terrane lacks volcanic rocks and contains a sequence of marbles, quartzites and quartzofeldspathic gneisses. These rocks were probably deposited between 1415 and 1300 Ma, and were intruded by plutonic rocks and

**subjected to metamorphism at roughly 1170 to 1160 Ma. The Frontenac Terrane is preserved at granulite facies, in contrast to the lower grade superterrane to the northwest, and the Adirondack Lowlands to the south. Mineral deposits of the superterrane consist of sulphides (pyrite and chalcopyrite predominate), carbonate-hosted zinc, magnetite skarns, gold in quartz-carbonate vein systems commonly associated with shear zones and a variety of industrial minerals. Mineral deposits in the Frontenac Terrane include graphite and other industrial minerals. A series of 1090 to 1070 million-year-old syenite intrusions occurs in all 3 terranes of the Central Metasedimentary Belt and may date the time when all 3 terranes began to behave as a single tectonic entity.**

**The Central Metasedimentary Belt Boundary Zone marks the boundary between the Central Gneiss Belt and the allochthonous terranes of the Central Metasedimentary Belt, which have been thrust onto the Central Gneiss Belt from the southeast. The boundary zone is a major shear zone several kilometres wide, characterized by strongly deformed rocks with northeasterly trending, moderately to shallowly southeast-dipping tectonic layering and southeast-plunging mineral lineations. It is subparallel with the Grenville Front Tectonic Zone in Ontario. Both zones have pronounced seismic expressions extending to the Moho. The Central Gneiss Belt may have served as basement to the supracrustal rocks of the Central Metasedimentary Belt, although there is growing evidence that the Central Gneiss Belt and the Central Metasedimentary Belt are allochthonous and only became attached to one another during the Grenville Orogeny.**

**Regional geochronology studies in the early 1960s indicated that the Grenville Province was characterized by *circa* 1000 to 950 million-year-old K-Ar biotite ages. These ages were attributed to postmetamorphic cooling coincident with the "Grenville Orogeny." It was initially believed that most structural and metamorphic features within the Grenville Province resulted from the *circa* 1.0 billion-year-old Grenville Orogeny. Subsequent Rb-Sr and U-Pb geochronologic studies showed that older rocks were present within the Grenville Orogen. Two orogenies are present within the Central Metasedimentary Belt, an older (ca. 1230 to 1180 Ma) "Elzevirian Orogeny" and a younger (ca. 1120 to 1050 Ma) "Ottawan Orogeny", the latter being equivalent to the "Grenville Orogeny." General consensus in the past has been that the younger orogeny was the more extensive of the two, although this view has been questioned recently. The tectonic evolution of the Grenville Province of Ontario, including the Elzevirian and Ottawan orogenies is synthesized in this chapter, and is outlined in time-sequential and cladogram approaches.**

**A variety of events affected the Grenville Province subsequent to the Ottawan Orogeny and prior to deposition of the Paleozoic sedimentary succession. These include: 1) injection of mafic dikes into the crust at *circa* 1140 to 900 Ma (Frontenac dike swarm), at *circa* 570 Ma (Grenville dike swarm) and at *circa* 420 Ma (Rideau dike swarm); and 2) alkalic and carbonatitic magmatism at *circa* 1050 to 1030 Ma immediately following the Ottawan Orogeny, and at *circa* 900, *circa* 570 and *circa* 180 Ma. Many of these dike swarms and alkalic events are related to development of, and latter reactivation along, the Ottawa–Bonnechere graben system. In addition, prior to deposition of the Paleozoic strata, the Grenville Province was peneplaned, and subjected to weathering producing paleosol horizons and karst.**

## INTRODUCTION

The Grenville Province was among the first parts of the Canadian Shield mapped in Canada (Logan 1863, 1866), and the first discovery of gold in Ontario was made near Eldorado in 1866. Considerable mapping was done in the Grenville Province in the late 19th century (cf., Adams and Barlow 1910, Vennor 1870), and study of the Grenville Province in Ontario contributed to many geologic concepts of the time (Easton 1989c). However, as the area became more populated, and significant mineral finds were made in northern Ontario, the economic *impetus* for geologic fieldwork in the Grenville Province diminished. It also soon became apparent that the geology of the Grenville Province was complex. This complexity discouraged many workers, and it is only recently that the study of Precambrian orogenic belts has advanced to the point that we can now begin to unravel the geology of the Grenville Province and to make meaningful comparisons with modern orogens.

The most complete recent summary of the Grenville Province in Ontario is that of Wynne-Edwards (1972). Sangster and Bourne (1982) and Carter (1984) provided overviews in their metallogenic studies of the Central Metasedimentary Belt in Ontario. Davidson (1984b, 1985, 1986a) summarized recent geologic developments in the Central Gneiss Belt in Ontario. The purpose of this chapter is to provide an overview of the Grenville Province in Ontario that reflects the considerable advances made over the last decade.

## REGIONAL SETTING OF THE GRENVILLE PROVINCE IN NORTH AMERICA

To understand the Grenville Province in Ontario, one must first place it in a regional tectonic setting. Major geologic regions in the midcontinent of North America are (Figure 19.1): the Southern Province, the Penokean Fold Belt, the Killarney Magmatic Belt, the Eastern Granite–Rhyolite Province, the Midcontinent Rift and the Grenville Province.

Sedimentary and minor volcanic rocks of the Huronian Supergroup of the Southern Province lie to the west (*see* Bennett et al., this volume; *see* Figure 19.1b). These rocks were folded and metamorphosed during the 1.8 billion-year-old Penokean Orogeny (Bickford et al. 1986; *see* Sutcliffe, this volume). The Huronian rocks southeast of Sudbury are separated from the Grenville Front by plutonic and minor volcanic rocks of the Killarney Magmatic Belt (van Breemen and Davidson 1988a), exposed along an 80 km stretch of the Grenville Front near Killarney (*see* Figure 19.1b). These rocks intrude the Huronian Supergroup (Davidson 1986b) and are 1750 to 1730 and 1470 to 1450 million years old (van Breemen and Davidson 1988a Southeast of the Grenville Front, Killarney Magmatic Bel rocks have been affected by the Grenville Orogeny. The Killarney granitoids have been interpreted (Easton 1986a; van Breemen and Davidson 1988a) to be an extension of one of several Mesoproterozoic orogenic belts underlying much of the North American midcontinent west of the area shown in Figure 19.1b (Bickford et al. 1986). The northern limit of the Grenville Province is considered, herein, to be the Grenville Front, consequently the Killarney Magmatic Belt north of the Grenville Front is not part of the Grenville Province (cf., Davidson 1986b, 1986c).

Felsic volcanic rocks and associated plutons of the Eastern Granite–Rhyolite Province form much of the basement beneath the midcontinental United States west of the Grenville Front in southern Michigan, Ohio and Indiana (*see* Figure 19.1b). These rocks are 1480 to 1420 and *circa* 1350 million years old (Bickford et al. 1986). Cutting this province is a northwest-trending belt of gravity and aeromagnetic anomalies (O'Hara 1981) interpreted to represent the buried portion of the 1.1 billion-year-old Midcontinent Rift (Keweenawan) (Hinze et al. 1982; *see* Sutcliffe, this volume).

Except near Manitoulin Island, basement rocks in southwestern Ontario have all been affected by a *circa* 1100 to 1070 million-year-old orogenic event, the "Grenville Orogeny." The Grenville Front (Derry 1950) marks the northwesterly limit of penetrative Grenvillian metamorphism and deformation and truncates structural trends in the adjacent structural provinces (*see* Figure 19.1), although locally some of these rocks can be traced across the Grenville Front for 10 to 30 km. As a result of recent work, it is now generally accepted that this orogenic event (or events) involved northwest-directed thrusting and imbrication of the entire crust, presumably as a result of a terminal collision at about 1100 Ma with a continental landmass somewhere to the southeast (cf., Moore 1986; Rivers et al. 1989). Many Rb-Sr and K-Ar ages in the Grenville Province, at about 1000 Ma, are synchronous with cooling after the "Grenville Orogeny." For the rocks exposed in the Grenville Province, this paper uses principal subdivisions as originally proposed by Wynne-Edwards (1972) with revisions (Moore 1982; Davidson et al. 1982; Easton 1988b, this chapter; Easton and Ford 1990a) (Figure 19.2).

## KILLARNEY MAGMATIC BELT

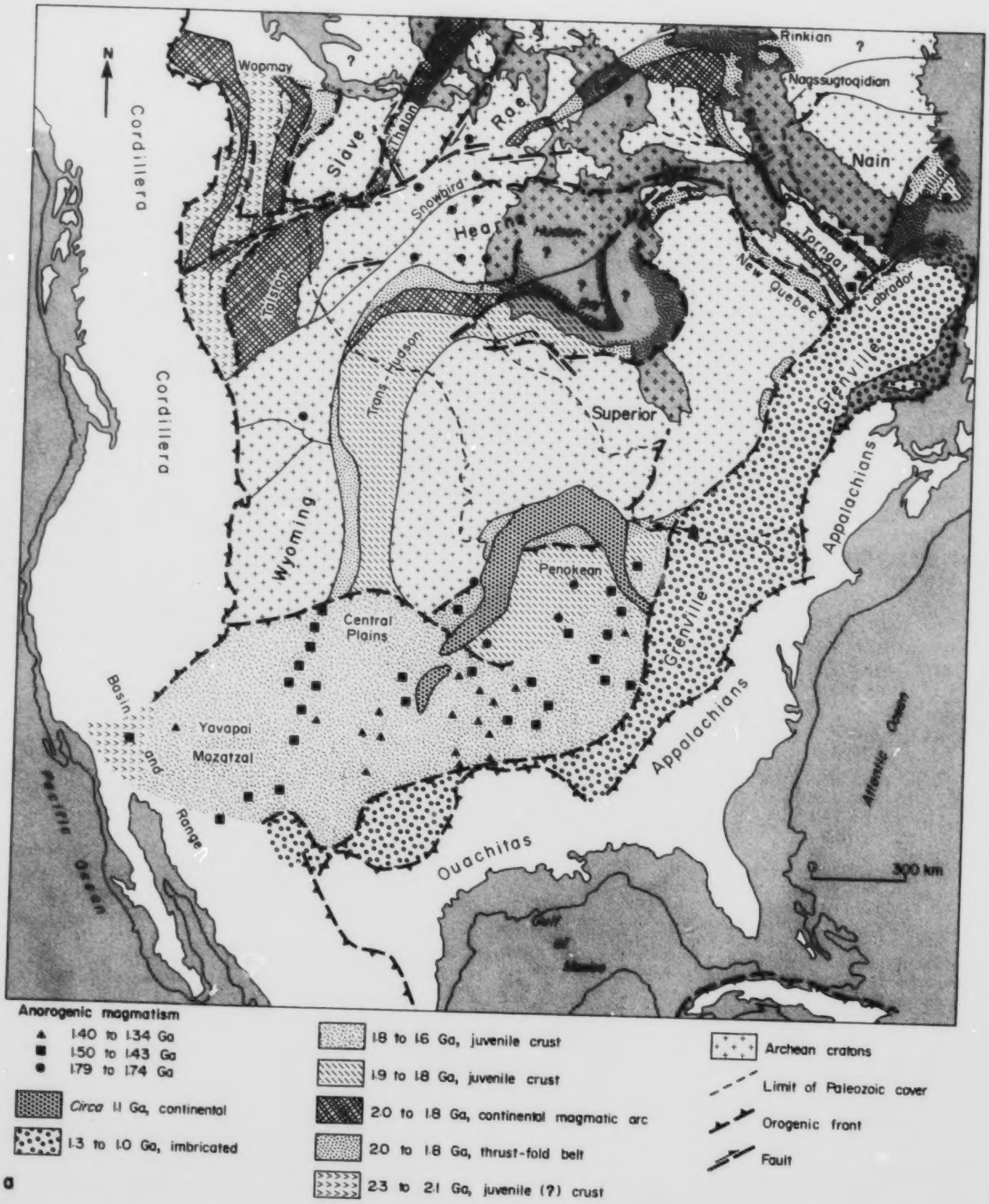
The Killarney Magmatic Belt consists of granitoid rocks, dated at 1750 to 1730 and 1550 to 1450 Ma (van Breemen and Davidson 1988a; Krogh 1989), with minor associated volcanic rocks. The belt is bounded to the north and west by metamorphosed rocks of the Huronian Supergroup, and to the south by the Grenville Front Boundary Fault (Figure 19.3). Metamorphosed equivalents of the Killarney Magmatic Belt, for example, the Johnnie Lake gneissic granite, have been recognized within the Grenville Front Tectonic Zone, and recent work suggests that the Killarney Magmatic Belt can be traced along the north shore of Lake Huron (Bethune 1989; McGrath et al. 1988; *see* Figure 19.3).

In the past, rocks of the Killarney Magmatic Belt have been variously assigned to both the Southern and the Grenville provinces. Quirke and Collins (1930), Brooks (1967, 1976), Stockwell (1969, 1982), Frarey and Cannon (1969) and Frarey (1985) assigned the Killarney rocks to the Grenville Province, using the Killarney–Huronian contact, regardless of whether the contact is intrusive or faulted, as the Grenville–Southern Province boundary. Ayres et al. (1971a), Card (1978), Card and Lumbers (1977), Lumbers (1975) and Davidson (1986b) assigned the Killarney rocks to the Southern Province, using the Grenville Front Boundary Fault, a mylonite zone which separates high-grade metamorphic rocks from lower grade metamorphic rocks (Figure 19.4), as the Grenville–Southern Province boundary. Easton (1986a) noted that the chronologic history of the Killarney Magmatic Belt differed from the rest of the northern Central Gneiss Belt, and Davidson (1986b, 1986c) presented convincing structural and geologic evidence indicating that the Killarney rocks were metamorphosed and deformed prior to the Grenville Orogeny, and that they represent a Mesoproterozoic magmatic belt distinct from the Penokean and Grenville orogens. Davidson (1986b) and Easton (1986a) both suggested that the Killarney Magmatic Belt could be traced to a belt of similar age rocks in the North American midcontinent. Additional mapping (Davidson and Bethune 1988; Culshaw et al. 1988) showed that Killarney rocks could be traced into the Grenville Front Tectonic Zone. Further, gravity data (McGrath et al. 1988; *see* Map 2595, map case) and regional aeromagnetic data (*see* Map 2587, map case) indicate that the Killarney Magmatic Belt is distinct, and that these geophysical characteristics can be traced 40 km west of the Grenville Front. Rivers et al. (1989) suggested the term Beaverstone Terrane for the Killarney equivalent rocks within the Grenville Province. This term is retained in this chapter as the Beaverstone Domain, which lies entirely within the Grenville Front Tectonic Zone.

Davidson (1986b, 1986c) characterized the sequence as a high-level volcano-plutonic complex. Preliminary geochemical studies of Killarney Magmatic Belt granitoids (Clifford 1990; Lumbers and Vertolli 1991) classify them as peraluminous, anorogenic granites.

Card (1976) first reported volcanic rocks from the Killarney area. Davidson (1986b, 1986c) confirmed the presence of felsic pyroclastic and volcaniclastic rocks (Figure 19.5), some of which are preserved as xenoliths and slivers within the Killarney Granite, and that the Lighthouse Porphyry may represent either a subvolcanic intrusion or a volcanic dome. Rocks associated with the Lighthouse Porphyry were interpreted by Davidson (1986b, 1986c) as silicified volcanic rocks, in contrast to earlier workers (Quirke and Collins 1930; Frarey 1985) who regarded them as feldspathized quartzites. Large rafts of Bar River quartz arenite of the Huronian Supergroup are also present within the Killarney Granite.

van Breemen and Davidson (1988a) dated the Killarney Granite at 1742±1.4 Ma and the Lighthouse Porphyry at $1732^{+7}_{-6}$ Ma. Samples of felsic volcanic rocks did not yield reliable results, yet they did clearly indicate that the volcanic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Pacific Ocean
Ouachitas
Gulf of Mexico
Atlantic Ocean
0
300 km
Anorogenic magmatism
▲ 1.40 to 1.34 Ga
■ 1.50 to 1.43 Ga
● 1.79 to 1.74 Ga
Circa 1.1 Ga, continental
1.3 to 1.0 Ga, imbricated
1.8 to 1.6 Ga, juvenile crust
1.9 to 1.8 Ga, juvenile crust
2.0 to 1.8 Ga, continental magmatic arc
2.0 to 1.8 Ga, thrust-fold belt
2.3 to 2.1 Ga, juvenile (?) crust
Archean cratons
Limit of Paleozoic cover
Orogenic front
Fault
a

**Figure 19.1a.** Principal subdivisions of the Precambrian of North America (using the terminology *of* Hoffman 1989) showing the extent of the Grenville Orogenic Belt.

rocks were not Huronian in age, and that they were younger than 1877 Ma (van Breemen and Davidson 1988a). The Bell Lake Granite (Figure 19.6) was also dated at 1471±3 Ma (van Breemen and Davidson 1988a), and Krogh et al. (1971) reported a preliminary age of 1665 Ma for the Chief Lake Granite. Pegmatites cutting the Killarney Granite that predate mylonitization along the front have been dated at 1400±50 Ma by U-Pb zircon (van Breemen and Davidson 1988a) and *circa* 1430 Ma by Rb-Sr muscovite (Krogh and Davis 1970b). Krogh (1989) also reports a U-Pb titanite age of 1453±7 Ma from leucosome pods in metasedimentary gneisses from the Killarney area, indicating a metamorphic event at *circa* 1450 Ma. Grenville metamorphism in the Beaverstone Domain has been dated by Krogh (1989) by zircon lower intercept data at *circa* 990 Ma. Titanite ages collected due east of the Grenville Front at Killarney show a discordant array, with an upper intercept of 1454±12 Ma and a lower intercept of 984±20 Ma, interpreted to represent metamorphic events at *circa* 1450 and 985 Ma (Haggart et al. 1992). The 1450 million-year-old event is coincident with magmatism and granulite-facies metamorphism in the Britt Domain to the southeast.

## DIVISIONS OF THE GRENVILLE PROVINCE IN ONTARIO

Over the past 30 years, several approaches have been used to subdivide the Grenville Province. The subdivision used here (*see* Figure 19.2) has been drawn from several previous subdivisions, in particular Wynne-Edwards (1972), Davidson et al. (1982), Davidson (1984b) and Moore (1982). Figure 19.7 summarizes these earlier schemes.

The Grenville Province in Ontario is separated into 2 major divisions (*see* bedrock map, Map 2544 and tectonic

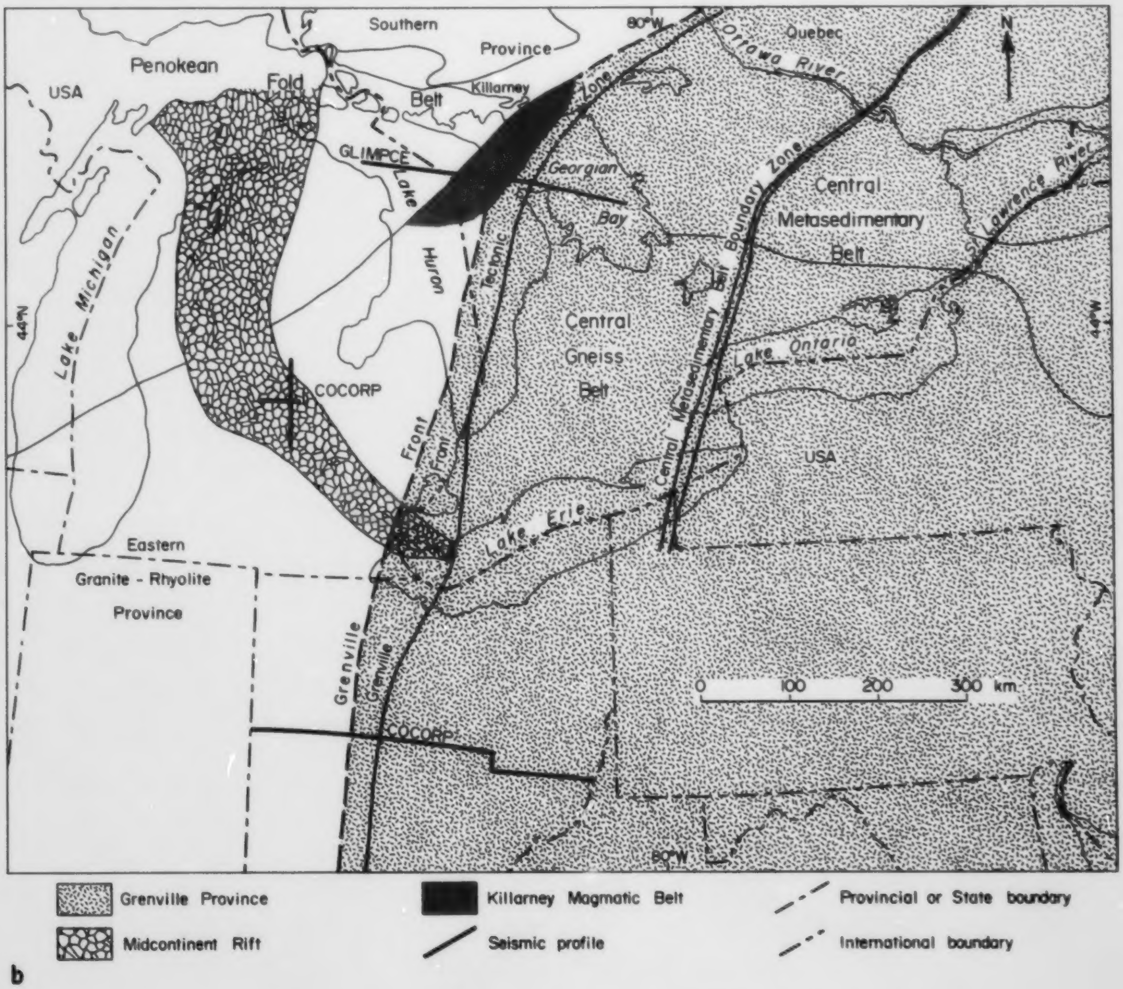
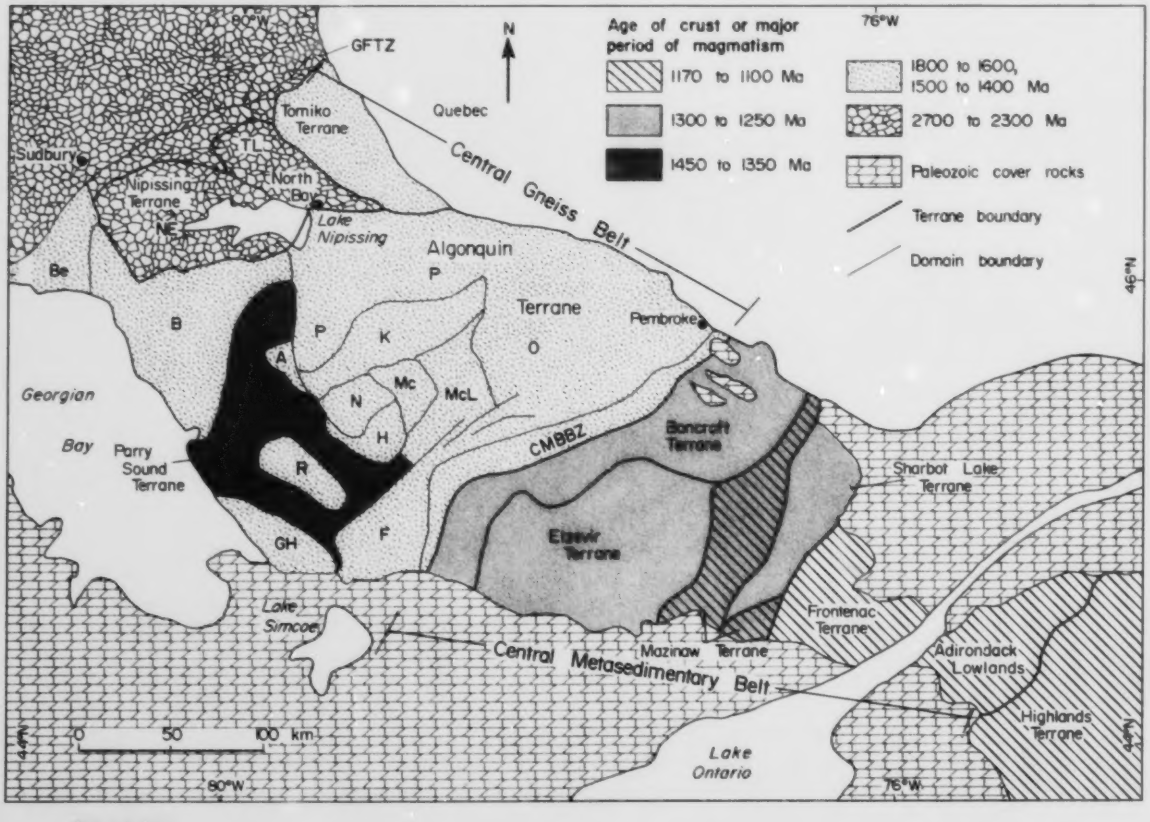

Southern
Province
Penokean
Fold
Belt
USA
Killarney
GLIMPCE
Lake
Huron
Georgian
Bay
80°W
Quebec
Ottawa River
N
Grenville Front Tectonic Zone
Central Metasedimentary Belt Boundary Zone
Central
Metasedimentary
Belt
St. Lawrence River
Central
Gneiss
Belt
Lake Ontario
Lake Michigan
44°N
44°N
COCORP
USA
Lake Erie
Eastern
Granite - Rhyolite
Province
Grenville Front
Grenville
0
100
200
300 km
COCORP
80°W
Grenville Province
Midcontinent Rift
Killarney Magmatic Belt
Seismic profile
Provincial or State boundary
International boundary
b

**Figure 19.1b.** Principal subdivisions of the Precambrian rocks in central and southern Ontario and adjacent areas, showing location of GLIMPCE and COCORP deep seismic profiles. Positions of the Grenville Front, Grenville Front Tectonic Zone, Central Metasedimentary Belt Boundary Zone and the southeastern extension of the Midcontinent Rift in the areas covered by Paleozoic rocks are interpretations *modified from* Carter and Easton (1990).

assemblage map, Map 2578, map case): the Central Gneiss Belt and the Central Metasedimentary Belt (Wynne-Edwards 1972). The Grenville Front (Derry 1950) separates the Grenville Province from older units to the north and northwest. That part of the Central Gneiss Belt that lies within 20 to 30 km of the Grenville Front has been termed the Grenville Front Tectonic Zone (cf., Wynne-Edwards 1972; Lumbers 1975, 1978; Rivers et al. 1989). The zone consists of deformed and metamorphosed rocks derived from the Southern and Superior provinces and the Killarney Magmatic Belt and is characterized by northeasterly trending zones of cataclasis, northeast-trending foliations and by a prominent southeast-plunging lineation (Lumbers 1975, 1978).

The Grenville Front Tectonic Zone (GFTZ), which includes the Grenville Front Boundary Fault, and the Central Metasedimentary Belt Boundary Zone (CMBBZ), which separates the Central Gneiss Belt from the Central Metasedimentary Belt (*see* Figures 19.2 and 19.7); are major shear zones several kilometres wide. These shear zones are characterized by strongly deformed rocks with northeasterly trending, moderately to shallowly southeast-dipping tectonic layering and southeast-plunging mineral lineations. Mylonites are locally developed. Marble tectonic breccia, containing fragments of metasedimentary, plutonic and amphibolite rocks is common within, and adjacent to, the eastern margin of the Central Metasedimentary Belt Boundary Zone. Significantly, the 2 zones are subparallel in Ontario (*see* Figures 19.1b and 19.2). Similar subparallel zones of intense deformation on a smaller scale form boundaries between lithotectonic terranes and domains within the Central Gneiss Belt and the Central Metasedimentary Belt (*see* Figure 19.2). The GFTZ and the CMBBZ both have pronounced seismic expressions


80°W
GFTZ
N
Age of crust or major period of magmatism
1170 to 1100 Ma
1300 to 1250 Ma
1450 to 1350 Ma
1800 to 1600, 1500 to 1400 Ma
2700 to 2300 Ma
Paleozoic cover rocks
Terrane boundary
Domain boundary
76°W
46°N
Tomiko Terrane
Quebec
Sudbury
TL
Nipissing Terrane
NE
North Bay
Lake Nipissing
Central Gneiss Belt
Algonquin
P
Terrane
O
Pembroke
Be
B
A
P
K
Mc
McL
N
H
R
Georgian Bay
Parry Sound Terrane
GH
F
CMBBZ
Bancroft Terrane
Elzevir Terrane
Sharbot Lake Terrane
Frontenac Terrane
Lake Simcoe
Central Metasedimentary Belt
Mazinaw Terrane
Adirondack Lowlands
Highlands Terrane
0 50 100 km
44°N
80°W
Lake Ontario
76°W
44°N

**Abbreviations**

| | |
|---|---|
| A | Ahmic Domain |
| B | Britt Domain |
| Be | Beaverstone Domain |
| CMBBZ | Central Metasedimentary Belt Boundary Zone |
| F | Fishog Domain |
| GFTZ | Grenville Front Tectonic Zone |
| GH | Go Home Domain |
| H | Huntsville Domain |
| K | Kiosk Domain |
| Mc | McCraney Domain |
| McL | McClintock Domain |
| MR | Moon River Domain |
| N | Novar Domain |
| NE | Nepewassi Domain |
| O | Opeongo Domain |
| P | Powassan Domain |
| PS | Parry Sound Domain |
| R | Rosseau Domain |
| S | Seguin Domain |
| TL | Tilden Lake Domain |

**Figure 19.2.** Lithotectonic terranes, domains and crustal ages within the Central Gneiss Belt and the Central Metasedimentary Belt in Ontario, based on Davidson (1984b, 1986a), Easton (1986a, 1988b), Moore (1982), Culshaw et al. (1983) and Dickin and McNutt (1989a, 1989b). Algonquin Terrane consists of the Ahmic, Britt, Fishog, Go Home, Huntsville, Kiosk, McCraney, McClintock, Novar, Opeongo, Powassan and Rosseau domains.

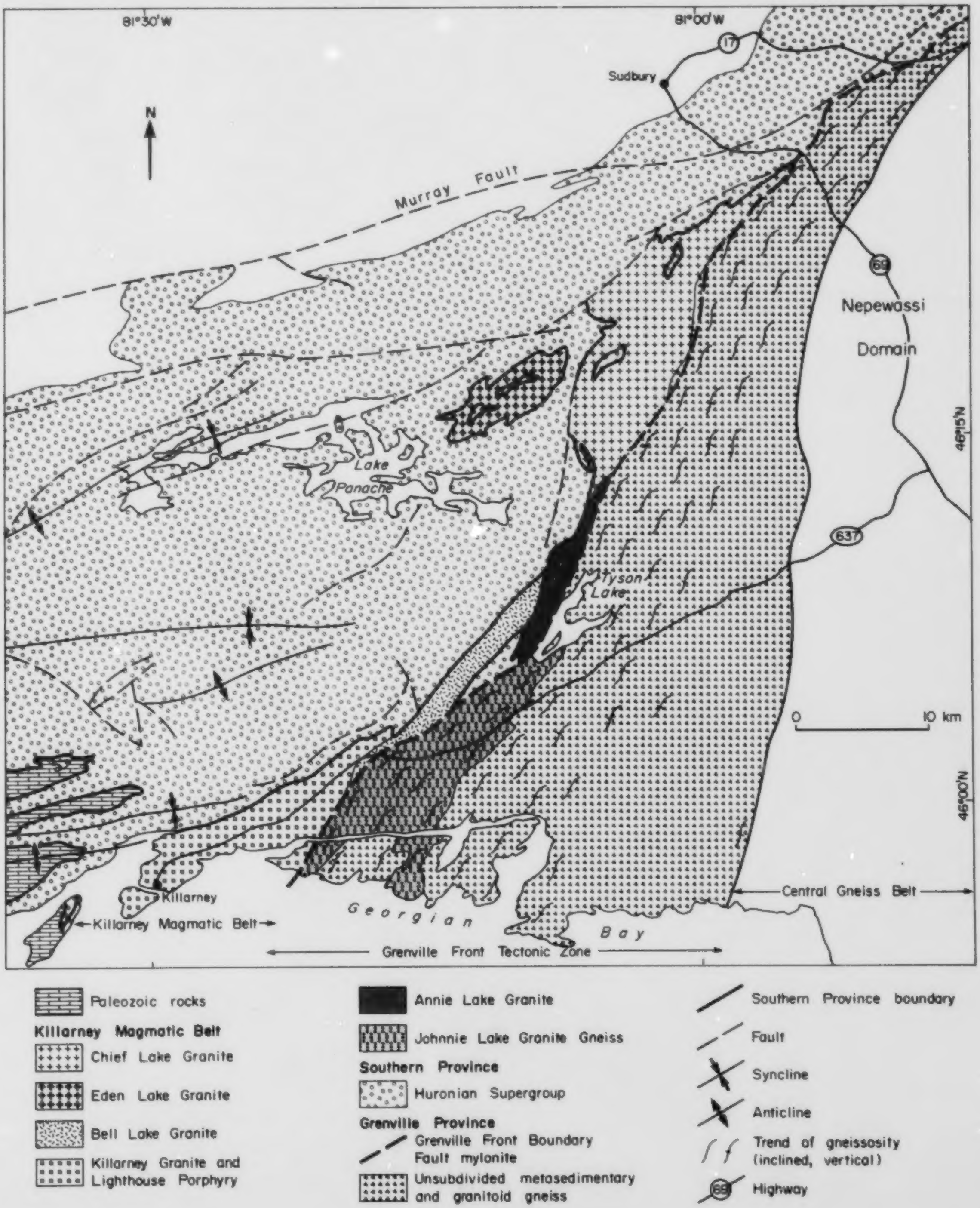

81°30'W
81°00'W
17
Sudbury
N
Murray Fault
69
Nepewassi
Domain
46°15'N
Lake
Panache
637
Tyson
Lake
0
10 km
46°00'N
Killarney
Killarney Magmatic Belt
Georgian
Bay
Central Gneiss Belt
Grenville Front Tectonic Zone
Paleozoic rocks
Killarney Magmatic Belt
Chief Lake Granite
Eden Lake Granite
Bell Lake Granite
Killarney Granite and Lighthouse Porphyry
Annie Lake Granite
Johnnie Lake Granite Gneiss
Southern Province
Huronian Supergroup
Grenville Province
Grenville Front Boundary Fault mylonite
Unsubdivided metasedimentary and granitoid gneiss
Southern Province boundary
Fault
Syncline
Anticline
Trend of gneissosity (inclined, vertical)
69
Highway

**Figure 19.3.** Simplified geologic map of the Killarney Magmatic Belt and surrounding geologic provinces.

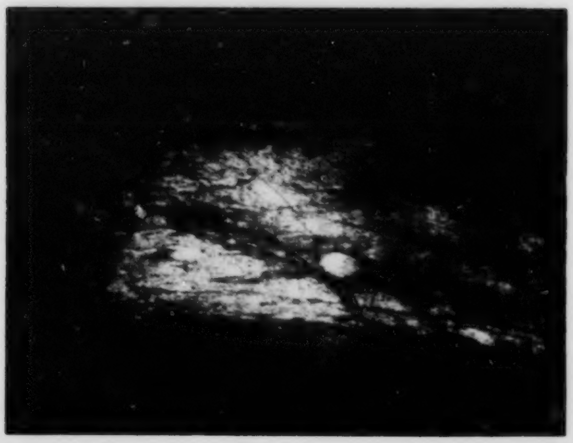


**Figure 19.4.** Close-up of deformed pegmatite within the Grenville Front Boundary Fault, southeast of Killarney. Hammer handle is 30 cm long.

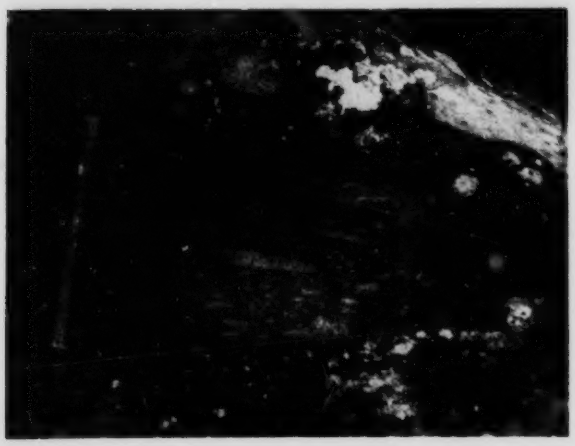


**Figure 19.5.** Felsic pyroclastic rocks from the Killarney Magmatic Belt. Hammer handle is 30 cm long.

(Figure 19.8) (Green et al. 1988, 1989; White et al. 1991a, 1992; Pratt et al. 1989; Culotta et al. 1989b) as discussed below.

## Central Gneiss Belt

The Central Gneiss Belt consists mainly of upper amphibolite- and local granulite-facies, quartzofeldspathic gneisses chiefly of igneous origin with subordinate paragneiss. In Ontario, the dominant structural trend is northeasterly, however, northwesterly trends occur along Georgian Bay (cf., Map 2587, map case). The Central Gneiss Belt consists of a variety of Archean to Mesoproterozoic crustal segments, all of which have been affected by the "Grenville Orogeny" (cf., Easton 1986a). As shown in Figure 19.2, rocks of 3 main ages are present (Easton 1986a; Dickin and McNutt 1989a, 1989b; Dickin et al. 1990). North of the French River, reworked Archean and Paleoproterozoic gneisses of the Nipissing Terrane (*see* Figure 19.2) are intruded by Mesoproterozoic (1700 to 1350 Ma) plutonic rocks, with granitic and monzonitic rocks predominant. The bulk of the Central Gneiss Belt (Algonquin and Tomiko terranes) consists of Mesoproterozoic gneisses (1800 to 1600 Ma) intruded by 1500 to 1400 million-year-old granitic and monzonitic plutons that may represent an extension of the Eastern Granite–Rhyolite Province (*see* Figure 19.1) across the Grenville Front. The Parry Sound Terrane (*see* Figure 19.2) consists of mafic to intermediate rocks extracted from the mantle at about 1450 to 1350 Ma (Dickin and McNutt 1989a).

**Figure 19.6.** Coarse-grained, potassium feldspar megacrystic granite of the Bell Lake Granite, Killarney Magmatic Belt. Key is 8 cm long.

Distinctive lithotectonic terranes, some further subdivided into domains, have been identified within the Central Gneiss Belt (terrane and domain terminology as outlined by Williams et al., this volume; Culshaw et al. 1983; Davidson 1984a, 1984b, 1986a; Easton 1988b) (*see* Figure 19.2). The terranes and domains are distinguished by differences in rock types, internal structure, metamorphic grade, geologic history and locally by geophysical signature. They are bounded by zones of intensely deformed layered rocks traceable for tens of kilometres (Davidson 1984a, 1986a; Rivers et al. 1989). Within the Grenville Front Tectonic Zone, the aeromagnetically distinct Beaverstone Domain (*see* Figure 19.2) is dominated by gneissic granitoid rocks probably derived from the Killarney Magmatic Belt. The Tomiko Terrane (formerly a domain, Easton 1988b) may consist entirely of Mesoproterozoic rocks that are allochthonous with respect to the adjacent Nipissing Terrane.

The Algonquin Terrane consists of quartzofeldspathic gneisses of plutonic and supracrustal origin characterized by a complex pattern of structural domains (Davidson 1986a, 1984a, 1984b; *see* Figure 19.2). It comprises the Britt, Algonquin, Kiosk and parts of the Muskoka domains of Davidson et al. (1982, 1985). Davidson et al. (1982) originally distinguished the Muskoka and Algonquin domains on the basis of structural trends and metamorphic grade. However, Nd-Sm model age data of Dickin and McNutt (1989a, 1990) and geochronologic evidence, combined with recent mapping, suggests the scheme used here, which relates areas of similar crustal and mantle-derivation ages. Thus, the Algonquin Terrane becomes an area of Mesoproterozoic, polycyclic rocks, consisting of a number of domains (domains and subdomains of Davidson et al. 1982, 1985). The southern and western parts of the Algonquin Terrane have been subdivided into the Britt, Ahmic, Kiosk, Rosseau, Go Home, Huntsville, Novar, McClintock (McLintock in Davidson et al. 1985 and subsequent publications), McCraney, Opeongo and Powassan domains, which are interpreted to represent the lowest portion of a stacked succession of thrust sheets in this region (Davidson 1986b). Large folded sheets of gneissic granites with primary isotopic ages in the 1500 to 1400 million year range (van Breemen et al. 1986; Easton 1986a) occur in all these domains. Rocks of this age are common in the Eastern Granite–Rhyolite Province (*see* Figure 19.1), and the Algonquin Terrane probably represents a section of Mesoproterozoic crust (1800 to 1600 Ma) extensively injected by granitic magmas of the Eastern Granite–Rhyolite Province. Although imbricated by later thrusting, the Algonquin Terrane is probably parautochthonous (Rivers et al. 1989).

Overlying and southeast of these predominantly parautochthonous terranes, throughout the Grenville Province in Canada, are several allochthonous terranes having no obvious lithologic counterparts with the terranes they overlie (Rivers et al. 1989; Figure 19.7g). Within the Central Gneiss Belt in Ontario, Rivers et al. (1989) identified only 1 allochthonous terrane, the Parry Sound Terrane. The Parry Sound Terrane is composed of 3 domains (*see* Figure 19.2). The lowermost and northernmost of these allochthonous thrust sheets is the Parry Sound Domain, composed of a distinctive assemblage of mafic to intermediate gneisses with subordinate marble, quartzitic and pelitic rocks that wrap around orthogneiss masses (Easton 1989d; Culshaw et al. 1989). Some workers have argued that volcanic rocks are present within the western Parry Sound Domain (Easton 1989d; Culshaw et al. 1989). Unlike other domains in the Central Gneiss Belt, rocks in the Parry Sound Domain have a steeply dipping, north to northwest-trending fabric. U-Pb ages between 1425 and 1350 Ma (van Breemen et al. 1986) from metaigneous rocks in the Parry Sound Domain, and Nd-model ages, indicate that the Parry Sound rocks were extracted from the mantle at about 1450 to 1400 Ma (Dickin and McNutt 1989a). Overlying the Parry Sound Domain are thrust sheets of layered or chaotic migmatite of the Moon River and Seguin domains.

The Fishog Domain of the Algonquin Terrane (part of an ill-defined region of the Muskoka Domain *of* Davidson et al. 1982) may be another allochthonous terrane. It is characterized by an intense aeromagnetic high, and tonalitic to dioritic gneisses intruded by magnetite-bearing granite and monzonite plutons (Easton 1988c, 1988g, 1990a). No isotopic ages are available from the Fishog Domain, consequently it is not known if it is indeed allochthonous (i.e., a terrane), or if it is part of the Algonquin Terrane (i.e., a domain).

## Central Metasedimentary Belt

The Central Metasedimentary Belt, southeast of the Central Metasedimentary Belt Boundary Zone, is a major Mesoproterozoic accumulation of marble, volcanic rocks and clastic metasedimentary rocks. The supracrustal rocks have been invaded by compositionally diverse, syntectonic, late tectonic and posttectonic plutonic rocks; and the entire succession has been metamorphosed at grades varying from greenschist to granulite facies. Most mineral deposits in the Grenville Province of Ontario are located within the Central Metasedimentary Belt. The Central Metasedimentary Belt can be divided into several lithotectonic terranes characterized by differences in rock type, geologic history, structural history, and ages of metamorphism and plutonism (*see* Figure 19.2).

The Bancroft Terrane consists mainly of deformed carbonate metasedimentary rocks, with minor volcanic rocks, and a distinctive suite of nepheline syenites and syenites intruded at *circa* 1270 to 1220 Ma (Miller 1983). Most of the Bancroft Terrane is at middle to upper amphibolite facies. Mineral deposits consist of carbonate-hosted zinc deposits, uranium mineralization associated with alkalic and perhaps carbonatitic rocks, molybdenum and a variety of industrial minerals.

The Elzevir Terrane, including the classical Grenville Supergroup, is characterized by volcanism and sedimentation between 1300 and 1250 Ma, followed by plutonism and metamorphism at 1250 to 1230 Ma and at 1130 to 1070 Ma. Large areas of the south-central and eastern Elzevir Terrane are preserved at greenschist facies. In those areas, primary sedimentary and volcanic textures are locally well-preserved. The Mazinaw Terrane consists of Grenville Supergroup marbles, calc-alkalic metavolcanic rocks and clastic metasedimentary rocks that were deposited on a basaltic-tonalitic basement, and which has a more intense and prolonged structural and metamorphic history that is similar to that present in the Frontenac Terrane (*see* below). In addition, Grenville Supergroup strata in the Mazinaw Terrane are overlain by a younger (ca. 1155 Ma), yet metamorphosed and deformed cover sequence, the Flinton Group, composed of quartz arenites, conglomerates and pelitic schists. It appears to be derived, in part, from weathering of Frontenac Terrane plutonic and metamorphic rocks (Kinsman and Parrish 1990).

The Sharbot Lake Terrane consists of a wedge of upper greenschist- to lower amphibolite-facies carbonate and mafic metavolcanic rocks located between the Mazinaw and Frontenac terranes. Although geographically separate from the Elzevir Terrane, the Sharbot Lake Terrane appears to be closely related to it in rock type and structural and metamorphic history. The Elzevir Terrane can be subdivided into several subdomains, each characterized by distinctive lithologic sequences and combinations of plutonic suites. These subdivisions of the Elzevir Terrane, and the Mazinaw and Sharbot Lake terranes, are discussed in greater detail elsewhere in this chapter. Mineral deposits of the Elzevir, Mazinaw and Sharbot Lake terranes consist of sulphides (pyrite and chalcopyrite predominate), carbonate-hosted

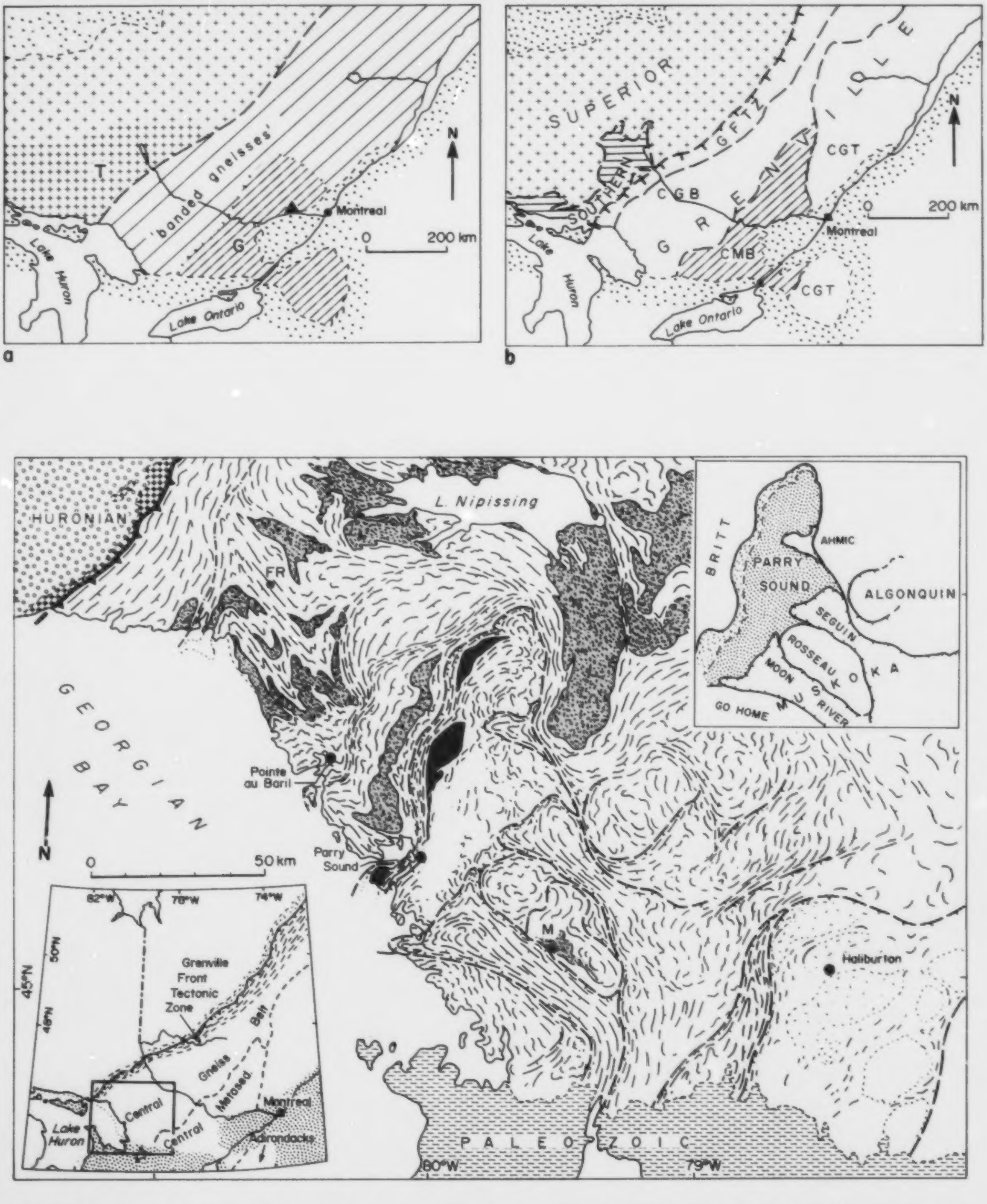

T
'banded gneisses'
G
Montreal
Lake Huron
Lake Ontario
N
0
200 km
a
SUPERIOR
SOUTHERN
GFTZ
GRENVILLE
CGT
CGB
CMB
Montreal
Lake Huron
Lake Ontario
N
0
200 km
CGT
b
HURONIAN
L. Nipissing
FR
GEORGIAN BAY
Pointe au Baril
Parry Sound
N
0
50 km
M
Haliburton
PALEOZOIC
80°W
79°W
BRITT
PARRY SOUND
AHMIC
ALGONQUIN
SEGUIN
ROSSEAU
MOON
GO HOME
MUSKOKA
RIVER
82°W
78°W
74°W
50°N
48°N
45°N
Grenville Front Tectonic Zone
Gneiss Belt
Central
Metased.
Central
Montreal
Adirondacks
Lake Huron

**Figure 19.7.** Various subdivisions of the Grenville Province from 1900 to present. **a)** Original concept of Grenville (G) and Timiskaming (T) subprovinces (bold patterns) *after* Wilson (1917a, 1917b, 1925) and subsequent expansion (light patterns) *after* Wilson (1939). Triangle notes Logan's initial description of the Grenville "series". Stipple indicates Paleozoic cover. *From* Davidson (1986a). **b)** Grenville Province subprovinces *after* Wynne-Edwards (1972), with structural provinces in the same area *modified from* Stockwell (1982). GFTZ—Grenville Front Tectonic Zone, CGB—Central Gneiss Belt, CMB—Central Metasedimentary Belt, CGT—Central Granulite Terrane. Stipple indicates Paleozoic cover. *From* Davidson (1986a). **c)** Domain subdivision of the Central Gneiss Belt in the Parry Sound–Huntsville area *from* van Breemen et al. 1986.

zinc, magnetite skarns, gold in quartz-carbonate vein systems commonly associated with shear zones and a variety of industrial minerals.

The Frontenac Terrane lacks volcanic rocks and contains a sequence of marbles, quartzites and quartzofeldspathic gneisses that were intruded by plutonic rocks and subjected to metamorphism at roughly 1170 to 1160 Ma. The Frontenac Terrane is preserved at granulite facies, in contrast to the lower grade Bancroft and Elzevir terranes to the north, and the Adirondack Lowlands to the south. Mineral deposits include graphite and other industrial minerals. A series of 1090 to 1070 million-year-old syenite intrusions occurs in all 3 terranes of the Central Metasedimentary Belt (Corriveau 1990) and may date the time when all 3 terranes began to behave as a single tectonic entity.

The Central Metasedimentary Belt Boundary Zone marks the boundary between the Central Gneiss Belt and the allochthonous terranes of the Central Metasedimentary Belt, which have been thrust onto the Central Gneiss Belt from the southeast. It is not known if, and to what extent, the Central Gneiss Belt served as basement to the supracrustal rocks of the Central Metasedimentary Belt, although there is growing evidence that the Central Gneiss Belt and the Central Metasedimentary Belt are allochthonous and only

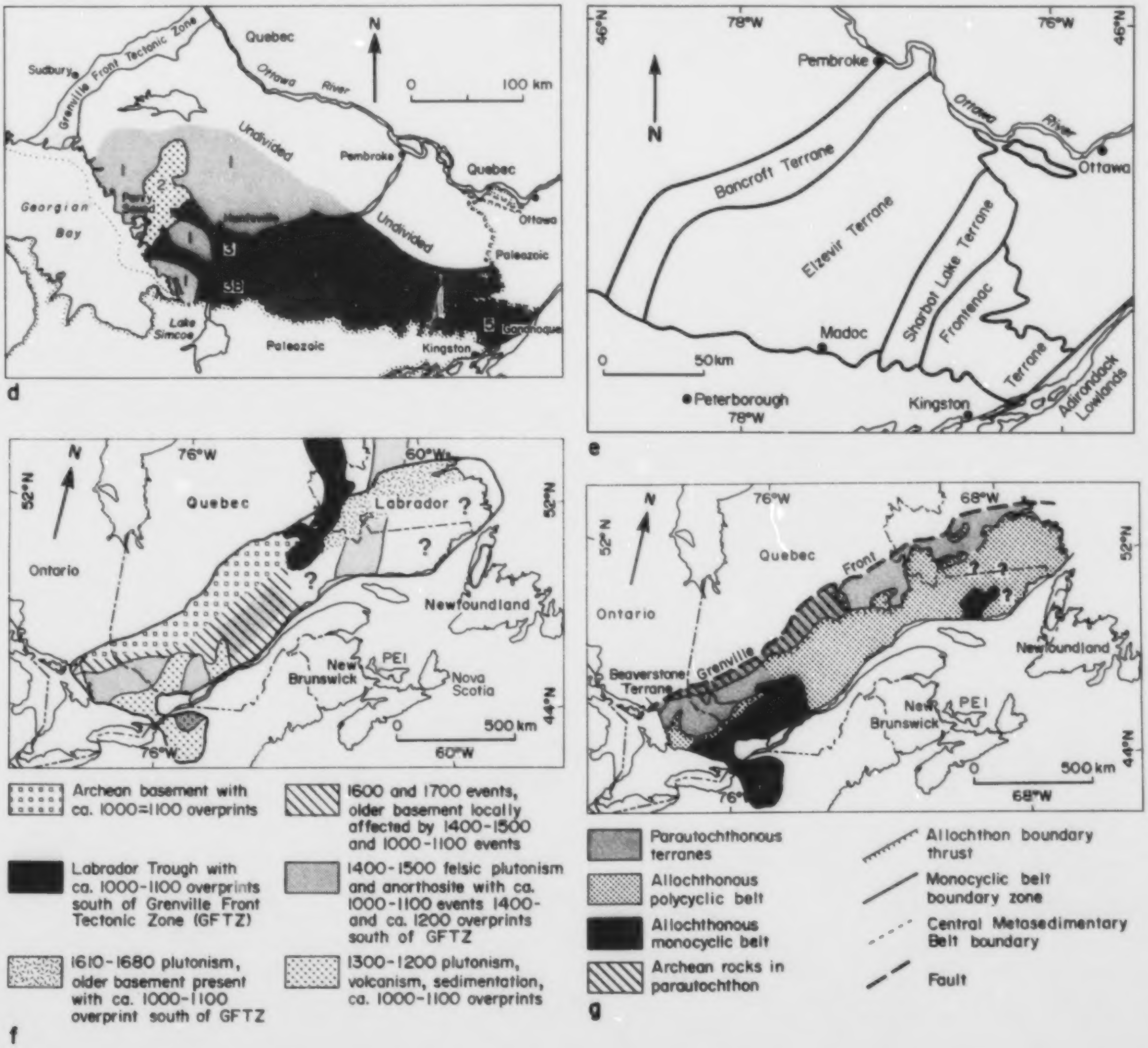
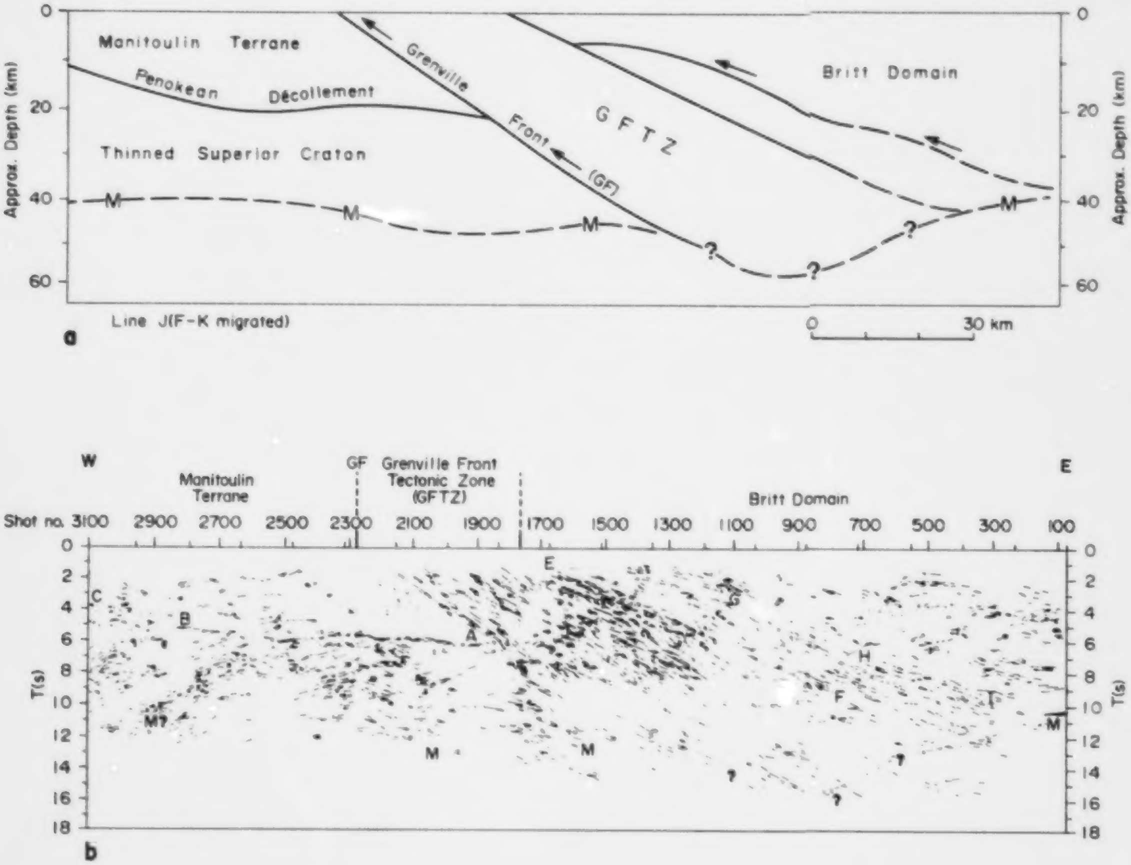

Grenville Front Tectonic Zone
Sudbury
Quebec
Ottawa River
N
0
100 km
Undivided
Pembroke
Quebec
Georgian Bay
Parry Sound
Ottawa
Paleozoic
Lake Simcoe
Paleozoic
Kingston
Gananoque
1
2
3
3B
5
d
46°N
78°W
76°W
46°N
Pembroke
Ottawa River
Ottawa
N
Bancroft Terrane
Elzevir Terrane
Sharbot Lake Terrane
Frontenac Terrane
Madoc
0
50km
Peterborough
78°W
Kingston
Adirondack Lowlands
e
N
76°W
60°W
52°N
Quebec
Labrador
?
Ontario
Newfoundland
New Brunswick
PEI
Nova Scotia
500 km
60°W
76°W
44°N
Archean basement with ca. 1000=1100 overprints
Labrador Trough with ca. 1000-1100 overprints south of Grenville Front Tectonic Zone (GFTZ)
1610-1680 plutonism, older basement present with ca. 1000-1100 overprint south of GFTZ
1600 and 1700 events, older basement locally affected by 1400-1500 and 1000-1100 events
1400-1500 felsic plutonism and anorthosite with ca. 1000-1100 events 1400- and ca. 1200 overprints south of GFTZ
1300-1200 plutonism, volcanism, sedimentation, ca. 1000-1100 overprints
f
N
76°W
68°W
52°N
Quebec
Front
Ontario
Beaverstone Terrane
Grenville
Newfoundland
New Brunswick
PEI
500 km
68°W
76°
44°N
Parautochthonous terranes
Allochthonous polycyclic belt
Allochthonous monocyclic belt
Archean rocks in parautochthon
Allochthon boundary thrust
Monocyclic belt boundary zone
Central Metasedimentary Belt boundary
Fault
g

**Figure 19.7.** Various subdivisions of the Grenville Province from 1900 to present. **d)** Structural stacking of domains and subdomains within the Central Gneiss Belt *after* Davidson (1984a, 1986c), Schau et al. (1986) and Easton (1986b, 1990a). **e)** Terrane subdivisions within the Central Metasedimentary Belt based on Moore (1982), Brock and Moore (1983) and Davidson (1986a). **f)** Chronologic provinces within the Grenville Province as outlined by Easton (1986a). **g)** Tectonic subdivision of the Grenville Province *modified from* Rivers et al. (1989).


Manitoulin Terrane
Penokean Décollement
Thinned Superior Craton
Grenville Front (GF)
GFTZ
Britt Domain
Approx. Depth (km)
0
20
40
60
M
?
Line J(F-K migrated)
0
30 km
a
W
E
GF
Grenville Front Tectonic Zone (GFTZ)
Manitoulin Terrane
Britt Domain
Shot no. 3100
2900
2700
2500
2300
2100
1900
1700
1500
1300
1100
900
700
500
300
100
T(s)
0
2
4
6
8
10
12
14
16
18
A
B
C
E
F
G
H
I
M
M?
?
b

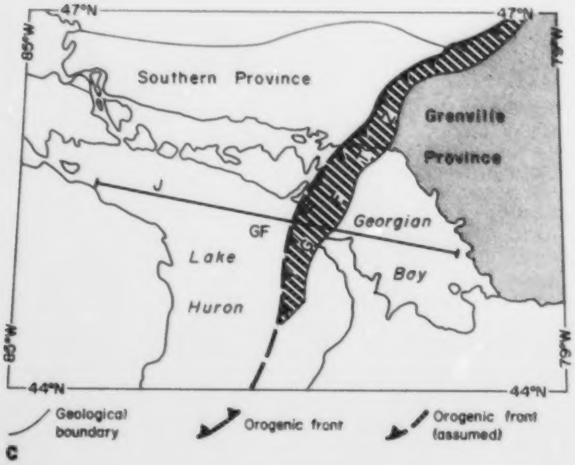

47°N
44°N
85°W
79°W
Southern Province
Grenville Province
Georgian Bay
Lake Huron
J
GF
Geological boundary
Orogenic front
Orogenic front (assumed)
c

**Figure 19.8.** Line diagram showing reflections recorded along GLIMPCE seismic profile. Figure only shows section from the eastern part of the line. Map (c) shows location of profile. Line diagram is based on F-K migrated section. Vertical scale is two-way traveltime in seconds. M is reflection Moho. Green et al. (1988) interpreted reflectors A-B-C as the master décollement of the 1890 to 1830 million-year-old Penokean Orogeny, with the upper crustal layer representing a thick wedge of deformed Huronian strata and displaced Archean basement or an allochthonous mass. Line E-G-H-I was interpreted by Green et al. (1988) as the northwestern edge of the Britt Domain and the GFTZ, the latter consisting of a number of crustal slices bounded by thrusts (e.g., area between Lines E-F and E-G-H-I). The Britt Domain contains fewer prominent reflectors than the GFTZ, although the dominance of eastward-dipping reflectors seems to be characteristic of the Grenville Province in the GLIMPCE profile. The cause of these prominent east-dipping reflectors is not known. Measurements of seismic velocity in Grenvillian ductile mylonites such as those within the Parry Sound shear zone show no difference from less-deformed protoliths from the Britt and Parry Sound domains (Davidson 1990b). *Modified from* Green et al. (1988, 1989).

became attached to one another during the Grenville Orogeny (cf., Davidson 1991).

# REGIONAL PATTERNS WITHIN THE GRENVILLE PROVINCE IN ONTARIO

Some aspects of Grenville Province geology are best discussed on a regional scale, because of limitations on the quality and quantity of data across the Grenville Province, and because regional contrasts provide a better picture of the distinctions between the lithotectonic divisions of the Grenville Province.

## Geophysics

### REFRACTION SEISMIC

The 1982 Canadian Consortium for Crustal Reconnaissance using Seismic Techniques (COCRUST) seismic refraction experiment (Mereu et al. 1986a, 1986b) ran profiles across parts of the Grenville Province (Figures 19.9 and 19.10). The experiments found no evidence of a mid-crustal discontinuity on any of the lines, however, the work showed pronounced velocity gradients in the upper crust, and that the Central Gneiss Belt is more homogeneous than the Central Metasedimentary Belt. They also found that the crust was thicker on the southeast side of the Central Metasedimentary Belt Boundary Zone (*see* Figure 19.10b) than beneath the Central Gneiss Belt, and suggested that the surface shear zones in the Central Metasedimentary Belt Boundary Zone may continue throughout the crust forming a depression on the Moho. If so, this structure was a zone of deformation, not a discrete fault. Preliminary data obtained from the Abitibi–Grenville LITHOPROBE seismic reflection line (White et al. 1991a, 1992) also suggests that the CMBBZ extends throughout the crust.

Southeast of Val d'Or, Quebec, the Grenville Front at depth is marked by a change in the character of the velocity gradient within the crust and by significant thickening (greater than 5 km) of the crust to the south of the Grenville Front (*see* Figure 19.10a; Mereu et al. 1986a, 1986b). Mereu et al. (1986a) suggest that the Superior Province was cooler

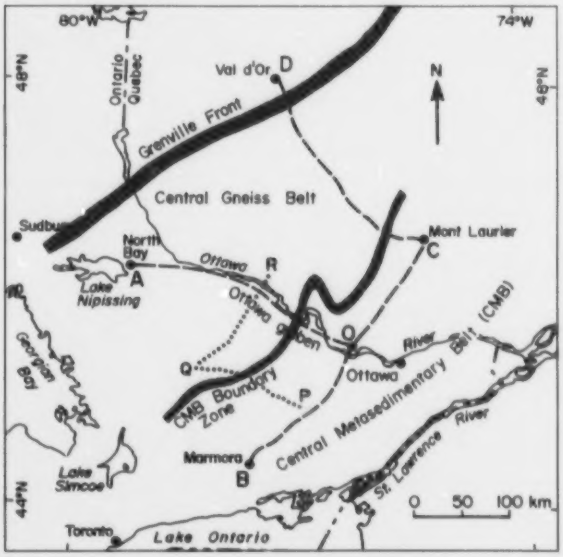


**Figure 19.9.** Map showing the location of the 1982 COCRUST experiment. Seismic refraction profiles are shown in Figure 19.10. Shot points A, B, O, C, D and lines AO, BO, OC and CD belong to the Ottawa-Bonnechere Graben experiment. Line QP indicates the location where the wide-angle reflected waves sampled the Moho, for the fan experiment, in which shot B was recorded along line AO. *Modified from* Mereu et al. 1986a.

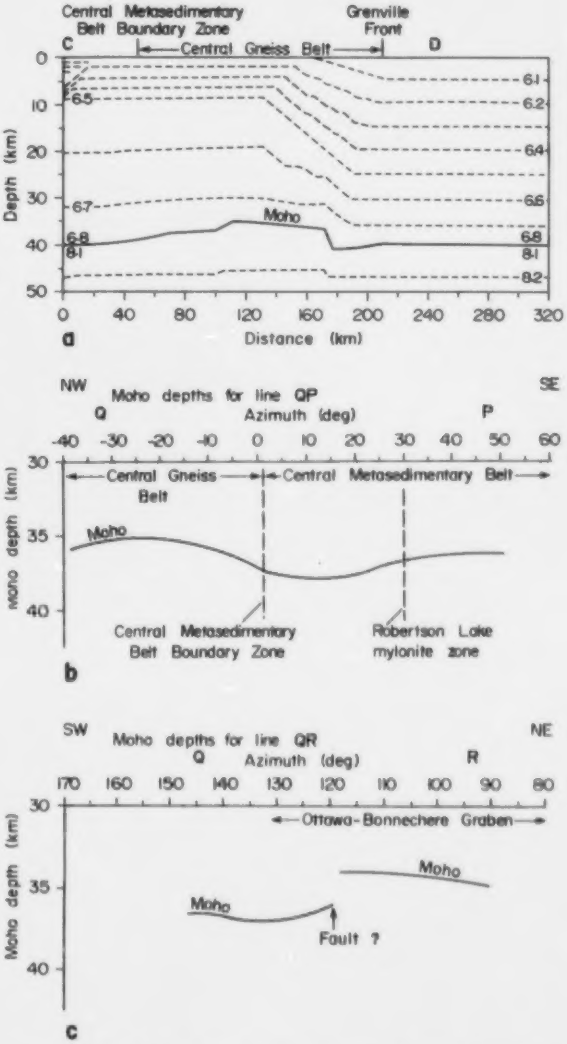


**Figure 19.10.** Seismic refraction profiles for the 1982 COCRUST seismic survey (*modified from* Mereu et al. 1986a). Location of lines are shown in Figure 19.9. Dashed lines in figure (a) are velocity contour lines showing velocities in km/s. a) Velocity gradient model for line CD. b) Moho depth profile along line QP across the Central Metasedimentary Belt Boundary Zone. c) Moho depth profile along line QR across the Ottawa–Bonnechere Graben.

and more rigid than the Grenville Province rocks to the south, and that northwest-directed thrusting formed the depression on the Moho in a similar fashion to the depression present along the Central Metasedimentary Belt Boundary Zone.

Mereu et al. (1986a, 1986b) report that the Moho along the Ottawa–Bonnechere Graben is poorly defined, possibly due to intrusion of upper mantle material into the lower crust, consistent with suggestions that the graben is an ancient rift (e.g., Kumarapeli and Saull 1966; Kumarapeli 1985).

## REFLECTION SEISMIC

In 1986, the Great Lakes International Multidisciplinary Program on Crustal Evolution (GLIMPCE) recorded 2 multichannel seismic reflection profiles in Lake Huron, one of which is shown in Figure 19.8 (Green et al. 1988, 1989; Epili and Mereu 1991). The area south of Manitoulin Island is characterized by curvilinear reflections, consistent with the complex geology of the region. The outstanding feature is a 32 km-wide band of east-dipping reflectors that coincide with the GFTZ (*see* Figure 19.8). Its steep western boundary delineates the Grenville Front Boundary Fault mylonite zone at depth and sharply truncates geologic and seismic structures to the west. Strong, east-dipping reflectors can be traced to greater than 14 s (Green et al. 1988), indicating that the GFTZ extends throughout the crust. The GFTZ seems to fan at depth, and at 9 s it is 50% wider than at the surface (Green et al. 1988). The Moho is ill-defined on the profiles, although Green et al. (1988) suspect that the Moho is quite deep on the Grenville side of the GFTZ, as indicated in seismic refraction experiments in Quebec (Mereu et al. 1986a, 1986b).

The interpretation of the various reflectors (Green et al. 1988) are shown in Figure 19.8. Similar east-dipping reflectors have been found in seismic profiles across the Grenville Front in southwestern Ontario (Easton and Carter 1991) and in Ohio (Culotta et al. 1989a, 1989b).

A seismic line in Lake Erie (D.A. Forsyth, Geological Survey of Canada, personal communication, 1990) shows a number of major crustal breaks which are roughly coincident with terranes mapped by drill core analysis and aeromagnetic interpretation by Carter and Easton (1990) outlined later in this chapter. In both the Ohio COCORP line and the unpublished Lake Erie line, east-dipping reflectors are only prominent to about 100 to 150 km east of the Grenville Front. Farther east, there are fewer reflectors, some are flat-dipping, others are east- or west-dipping. A zone of west-dipping reflectors occurs west of the southward extension of the Central Metasedimentary Belt Boundary Zone. This zone of reflectors is located near the terminus of both reflection lines and is not adequately profiled.

As part of the Abitibi–Grenville LITHOPROBE Project, seismic reflection lines were run across the Parry Sound Terrane along highways 124 and 520 and the Central Metasedimentary Belt Boundary Zone and the Elzevir Terrane along Highway 62 from Dwight to Napanee *via* Tweed in the summer of 1991. Preliminary results of the southern line were available as this volume went to press (White et al. 1991a, 1991b, 1992). Most significant were the observations that: 1) the CMBBZ and the Bancroft Terrane appeared to be a single seismic entity; 2) the CMBBZ and Bancroft Terrane zone of shallow, southeast-dipping reflectors extends to the base of the crust, similar to the GFTZ; and 3) major seismic boundaries were present coincident with the terrane boundaries proposed herein (e.g., Elzevir–Mazinaw boundary).

## AEROMAGNETIC DATA

Ontario Geological Survey Map 2587, the shaded image of total magnetic field map of southern Ontario, uses a sun angle chosen to emphasize Precambrian geologic structure. Within the Central Gneiss Belt, strong magnetic highs are associated with: 1) the Killarney Magmatic Belt and its extension into the GFTZ; 2) the Parry Sound Domain; 3) the Fishog Domain which can be traced in subsurface to Lake Erie (Carter and Easton 1990); 4) a cluster of Neoproterozoic to Early Paleozoic alkalic and carbonatite complexes near Lake Nipissing; 5) the northern part of the Tilden Lake Domain and the River Valley–Temagami Segment of the GFTZ, perhaps associated with Paleoproterozoic mafic rocks at surface and at depth related to the *circa* 2480 million-year-old gabbroic anorthosite intrusions, such as the River Valley Anorthosite, and the *circa* 2220 million-year-old Nipissing Diabase Suite; and 6) the Cosby Granite and the French River area of the southern Nepewassi Domain. This may suggest that the Nepewassi Domain should be separated into 2 domains. The Grenville Front and the Grenville Front Tectonic Zone do not stand out prominently, in contrast to Quebec, where the front is defined by a chain of elongate, northeast-trending positive anomalies, with the GFTZ being a zone of little local magnetic relief (Wynne-Edwards 1972; Rivers et al. 1989). This difference in magnetic signature probably reflects the limited extent of Archean basement in the Grenville Province in Ontario. In the past, the magnetic high associated with the Killarney Magmatic Belt has been used to trace the Grenville Front south into Michigan and Ohio. As noted earlier, gravity and geochronologic data must be used in conjunction with aeromagnetic information in tracing the Grenville Front southward, as the Grenville Front cuts across the Killarney Magmatic Belt.

Additional features prominent on the aeromagnetic map in the Central Gneiss Belt include: 1) the regional structural trends within the Moon River and Rosseau domains; 2) the east-striking faults and dikes of the Ottawa–Bonnechere graben system and the Grenville dike swarm that cut through the Nipissing and Algonquin terranes at roughly 46°05′N; 3) a relatively flat magnetic pattern associated with the thickest accumulation of supracrustal rocks within the Tomiko Terrane; and 4) a subdued magnetic pattern in the northern Nepewassi Domain and the Grenville Front Tectonic Zone near Sudbury, similar to the subdued magnetic pattern associated with Archean-age rocks south of the Grenville Front in Quebec. Although the Central Metasedimentary Belt Boundary Zone shows prominently in the south where it has a northerly trend, in part, because of the presence of ferromagnetic Fishog Domain rocks to the west, the Central Metasedimentary Belt Boundary Zone is not magnetically prominent where it trends northeasterly from Haliburton to Pembroke.

Conspicuous features of the Central Metasedimentary Belt on the aeromagnetic map include: 1) the domal structures along the Harvey–Cardiff Arch; 2) a prominent magnetic high along the eastern flank of the Harvey–Cardiff Arch, coincident with the Burleigh fault, linking with the McArthurs Mills shear zone near Bancroft; 3) a flat magnetic pattern underlain by the Weslemkoon and Elzevir tonalites and their surrounding volcaniclastic aprons; 4) an alternating pattern of east-northeast-striking highs and lows associated with the supracrustal rocks and tonalites of the Mazinaw Terrane; 5) magnetic lows and subdued magnetic textures associated with carbonate basins in the Bancroft and the western Elzevir terranes; 6) localized, circular magnetic highs in the Elzevir and Frontenac terranes associated with *circa* 1090 to 1075 million-year-old Skootamatta Suite plutons. Similar circular highs beneath the Paleozoic cover underneath Lake Ontario may represent additional plutons of the Skootamatta Suite. Alternatively, they could have no exposed surface counterpart; and 7) the intense, curvilinear pattern characteristic of the granulite-facies rocks of the Frontenac Terrane.

## GRAVITY DATA

### Regional Gravity Trends

In Quebec, the Grenville Front Tectonic Zone is marked by a major linear negative Bouguer anomaly with values of about -100 mGal (Wynne-Edwards 1972; Rivers et al. 1989), probably the result of crustal thickening on the Grenville side of the front (Wynne-Edwards 1972). Within Ontario, this anomaly is not as prominent (*see* Maps 2594 and 2595, map case). A Bouguer gravity low along the Grenville Front in the Killarney area is associated with the Killarney Magmatic Belt and not the Grenville Front (McGrath et al. 1988; Figure 19.11; *see* Map 2595, map case). Within the remainder of the Grenville Province in Ontario, the most notable Bouguer anomalies are: 1) a prominent high associated with the Parry Sound Domain (*see* Figure 19.11; *see* Map 2595, map case); 2) a prominent high associated with the Nepewassi Domain, particularly the northwestern half of the domain (*see* Map 2595, map case); and 3) a prominent, east-trending low between North Bay and Mattawa (*see* Map 2595, map case) which does not correlate with any single geologic feature.

Real and Thomas (1987) note a steep gravity gradient belt (western boundary gradient belt) that trends northeasterly, roughly coincident with the Central Metasedimentary Belt Boundary Zone (CMBBZ), particularly in western Quebec and from the Ottawa River to Bancroft. South of Bancroft, the anomaly trends south along the axis of the Harvey–Cardiff Arch, at least 50 km east of the Central Metasedimentary Belt Boundary Zone and the prominent magnetic anomaly associated with the southward-trending segment of the Central Metasedimentary Belt Boundary Zone.

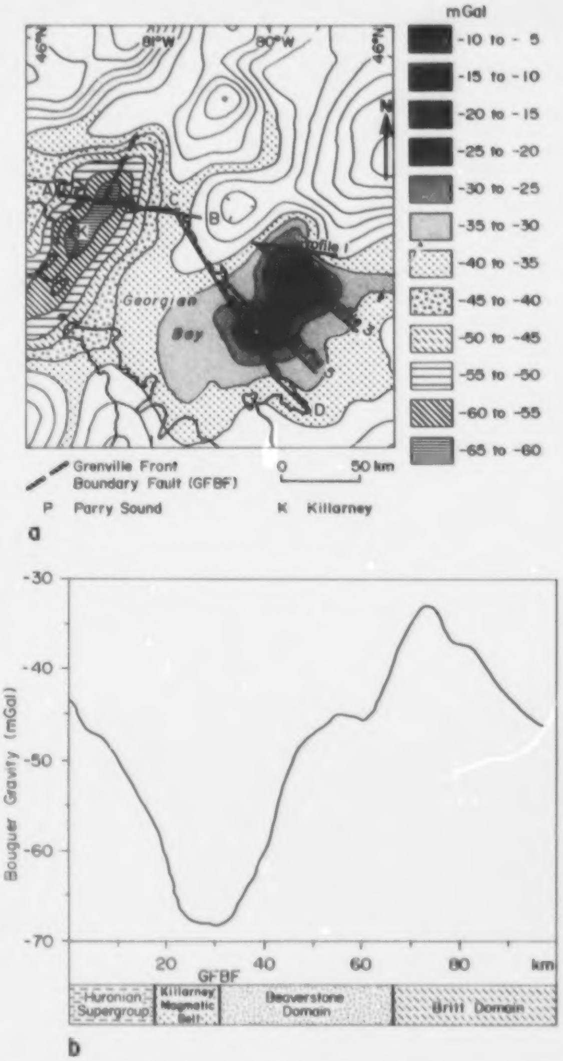


**Figure 19.11.** a) Detailed Bouguer gravity map of the Killarney area (*modified from* McGrath et al. 1988). b) Gravity profile along section A-B across the Killarney Magmatic Belt and the Grenville Front Boundary Fault. Location of section shown above (*modified from* McGrath et al. 1988). Gravity profiles 1,3 and 5 shown in Figure 19.12.

### Detailed Gravity Surveys

McGrath et al. (1988) interpreted a low near Killarney (*see* Figure 19.11) as the result of less dense Killarney Magmatic Belt felsic plutonic rocks, and concluded that Killarney Magmatic Belt rocks occur at least 12 km east of the Grenville Front Boundary Fault. This is consistent with field work by Davidson and Bethune (1988) who have traced felsic gneisses equivalent to the Killarney Magmatic Belt at least 14 km west of the Grenville Front Boundary Fault.

The Parry Sound Terrane is characterized by a 35 mGal positive gravity high. Lindia et al. (1983) and Lindia (1983) modelled the Parry Sound Domain with a bowl-shaped geometry, with depths to the base of the denser substrate ranging from 5 km in the northeastern and southwestern sections of the Parry Sound Domain to a maximum of 13 km in the central region (Figure 19.12). In addition, denser rocks could be traced eastward as tapering wedges beneath the Moon River and Sequin domains. The gravity data is consistent with northwest-directed thrusting for the Parry Sound region (Davidson et al. 1982; Culshaw et al. 1983); after thrust emplacement of the Parry Sound Domain, the great mass of dense rocks caused downwarping. An additional profile (McGrath et al. 1988) across the Parry Sound Domain indicated that the domain extends at least 5 to 10 km into Georgian Bay.

Real and Thomas (1987) modelled 6 profiles across the Elzevir and Mazinaw terranes to see if gravity data could be used to trace the Central Metasedimentary Belt Boundary Zone beneath the Central Metasedimentary Belt. They found that most of the gravity variation could be accounted for by structures within the uppermost 4 km of the crust. Their study was inconclusive regarding the existence of a décollement underlying the 4 km thick surface skin. Their model suggests that felsic intrusions in the area had tabular or sheet-like geometries that fell into 2 thickness categories: 1) the Weslemkoon, Elzevir and Northbrook tonalites and the Abinger Granite (2.4 to 4.2 km thick); and 2) the Cross Lake and Mellon Lake tonalite to granodiorite plutons and the Addington and Deloro granites (0.2 to 1 km thick). It is notable that all the thin sheets indicated by the model occur within the Mazinaw Terrane. Thus, the gravity model may reflect variations related to the different tectonic divisions within the Central Metasedimentary Belt.

## AIRBORNE GAMMA-RAY SPECTROMETRIC DATA

### Regional Patterns

Most of the Grenville Province is covered by 1:250 000 airborne gamma-ray spectrometric coverage (OGS 1978a-e), and much of the Central Metasedimentary Belt has been mapped at 1:50 000 scale (Ford et al. 1992; Geological Survey of Canada, unpublished data). The 1:250 000 scale gamma-ray data outlines much of the large-scale lithologic variation present within the Grenville Province in Ontario. The airborne gamma-ray maps reflect the broad crustal divisions of the Central Gneiss Belt, for example, the Parry Sound Domain and Nipissing Terrane both have very low abundances of uranium, thorium and potassium. The Tomiko Terrane is similar to the Algonquin Terrane in its gamma-ray signature, having higher levels of potassium, thorium and uranium than the adjacent Tilden Lake Domain. The Killarney Magmatic Belt, in particular the area underlain by the Chief Lake Granite, is high in uranium and thorium in comparison to adjacent regions. The Fishog Domain, coincident with an aeromagnetic high, also shows higher potassium and total count values than adjacent regions, due to well-exposed syenogranite intrusions (Easton 1986b, 1988c, 1988g, 1990a).

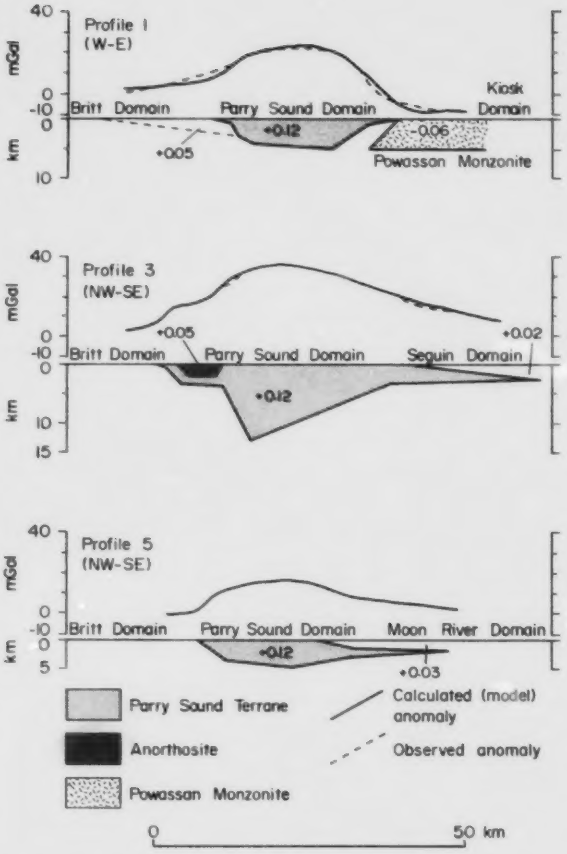


**Figure 19.12.** Gravity profiles across the Parry Sound Terrane showing the thin-skinned nature of the dense Parry Sound Terrane rocks (*modified from* Lindia et al. 1983, only 3 of the 5 profiles are illustrated). Location of profiles shown on Figure 19.11a, profiles 1, 3, 5. Density contrasts of bodies are in g/cm$^3$.

Similarly, large-scale tectonic subdivisions of the Central Metasedimentary Belt can be seen on the airborne gamma-ray spectrometric maps, particularly where 1:50 000 scale coverage exists. In part, the sensitivity of the gamma-ray spectrometric data to terrane boundaries is probably a reflection of the wide variety of rock types present in the Grenville Province with characteristic signatures.

### Uranium and Thorium Mineralization

As noted in Easton (1986b, 1990a, 1990f), there is excellent agreement between airborne gamma-ray spectrometric anomalies and uranium-thorium mineralization within the Grenville Province. Mineralization is concentrated mainly within several areas of the Central Metasedimentary Belt, with the most highly mineralized areas located near major regional structures, such as the CMBBZ, the flanks of the Harvey-Cardiff Arch and along major regional structures within the Kaladar region. This close agreement makes it unlikely that significant undiscovered uranium-thorium deposits are present in the Grenville Province in Ontario.

## PALEOMAGNETISM

As paleopoles were acquired from an increasing number of Grenville rocks (Hargraves and Burt 1967; Buchan and Dunlop 1973; Palmer and Carmichael 1973; Irving, Park and Emslie 1974), it became clear that the Grenville poles did not fall on the Apparent Pole Wander Path (APWP) for the period 1300 to 1100 Ma which had been determined from unmetamorphosed rocks of interior North America (Interior Laurentia). It was therefore suggested that the paleomagnetic data recorded the convergence and collision of a "Grenville Plate (Grenvillia)" with Laurentia in that time interval (Irving, Emslie and Ueno 1974; Buchan and Dunlop 1976). The convergent loop on the APW path was referred to as the "Grenville Loop", and this scenario required a suture somewhere in the central Grenville Province. With the advent of argon-argon thermochronometric studies (Berger et al. 1979; Berger and York 1981a, 1981b; *see* review *in* York 1984), it became clear that the Grenville paleopoles were much younger than previously suspected (950 to 850 Ma). Locally, some poles were even younger, being affected by a *circa* 450 million-year-old resetting (Lopez-Martinez and York 1983; York 1984). Thus, if a collisional suture is present in the Grenville Province in Ontario, its existence can not be verified by paleomagnetism.

Paleomagnetic studies have been used to determine the extent of Grenvillian metamorphic overprinting north of the Grenville Front. Hyodo et al. (1986) and Stupavsky and Symons (1982a) found little evidence of Grenvillian paleomagnetic resetting more than 2 to 5 km north of the Grenville Front in the Temagami and River Valley areas, respectively. Stupavsky and Symons (1982a) demonstrated that folding in the foreland region north of the front was pre-Grenvillian in age. Stupavsky and Symons (1982b), in a study of the Pickerel Complex and the Servos Tonalite south of Sudbury, reported remanent magnetizations lying on APW paths at 1725 and 1250 Ma. They attributed the former to time of emplacement, and the latter to emplacement of Sudbury Swarm diabase dikes. Both ages are coincident with known Proterozoic events in the region (*see* Central Gneiss Belt, below; Easton 1992), and it is possible that Grenvillian metamorphism did not completely reset magnetites in the area at 985 Ma.

Summary papers on Grenville paleomagnetic studies not listed above include Symons (1978), Irving and Lapointe (1975), Irving (1979), Roy and Robertson (1979), Irving and McGlynn (1981) and Park and Emslie (1983). Additional studies of paleomagnetism in the Central Metasedimentary Belt of Ontario include Dunlop et al. (1980), Dunlop and Schutts (1983) and Dunlop and Stirling (1985).

# Geochemistry

## REGIONAL LAKE SEDIMENT SURVEYS

Regional lake sediment geochemical surveys are only available for the southeastern part of the Grenville Province (OGS–GSC 1977a, 1977b; Hornbrook et al. 1984a, 1984b), but they do show regional patterns related to lithotectonic divisions of the area. For instance, the Sharbot Lake Terrane has a higher background level of lead, zinc, uranium and cobalt. The former 3 elements are probably related to the abundance of low-metamorphic grade carbonate metasedimentary rocks, with the last element reflecting the abundance of gabbro intrusions in the terrane. Arsenic shows a strong association with major regional structures, including the Elzevir–Mazinaw terrane boundary, the Robertson Lake mylonite zone and major faults associated with the Flinton Group in the Fernleigh–Ompah area of the Mazinaw Terrane. In addition to being enriched in carbonate terranes, zinc is also associated with calc-alkalic volcanic rocks of the Belmont Lake and Northbrook–Cloyne areas. Lead and zinc anomalies are also associated with the Ottawa–Bonnechere graben system. Easton and Ford (1990a) reported lead anomalies along faults of this system in the Grimsthorpe Domain, and Paleozoic-age, lead-carbonate vein systems in the southern Grenville Province are known to be associated with younger fault structures (Carter 1984; Kumarapeli 1976; Sangster and Bourne 1982).

## REGIONAL TILL AND OTHER SURFICIAL MATERIAL SURVEYS

The Geological Survey of Canada has amassed geochemical data for till and other glacial sediments across the southern Grenville Province (Kettles and Shilts 1983, 1989; Kettles 1989, 1990). The till geochemistry dataset for the eastern half of NTS 31C and 31F (Kettles 1990) shows similar patterns to the lake sediment data (OGS–GSC 1977a, 1977b).

The Sharbot Lake Terrane has higher than average concentrations of zinc, cadmium, lead and uranium, cobalt, chromium, copper and nickel compared to the Frontenac Terrane. The first group of elements are related to the carbonate-rich nature of the Sharbot Lake Terrane, whereas the second group of elements is probably related to gabbroic intrusions (e.g., Lavant Gabbro and Mountain Grove Gabbro). Gold, arsenic, mercury and silver are associated with the Robertson Lake mylonite zone (Easton 1988a, 1988f), a major regional structure with known copper-gold-antimony occurrences. Arsenic, gold, zinc and mercury are also elevated in the Fernleigh–Ompah area, in association with the Flinton Group and a set of northeast-trending lithology-parallel faults and shear zones.

A variety of studies in the southern Central Metasedimentary Belt have shown that dispersal trails in till associated with mineral deposits only extend a few kilometres from the deposit (e.g., DiLabio et al. 1982; Gleeson et al. 1984, 1986, 1989; Rampton et al. 1986). Anomalies in the regional till database are thereby little removed from their sources, and the regional data can be used to pinpoint areas for exploration.

# Metamorphism

## OVERVIEW OF "GRENVILLIAN" METAMORPHISM IN ONTARIO

### Pressure-Temperature Conditions

Figure 19.13 summarizes the regional variation in metamorphic conditions across the Grenville Province. Figure 19.14 is a compilation map showing the timing of metamorphism across the Grenville Province, based on U-Pb zircon, titanite and monazite ages. Most of the regional metamorphic data from the Central Gneiss Belt is from the work of Anovitz and Essene (1990). Detailed studies are listed in Table 19.1. Variation in peak metamorphic temperatures (*see* Figure 19.13a; Table 19.2) shows little correlation with known large-scale tectonic divisions, however, peak pressures (*see* Figure 19.13b) show a greater correlation; in particular, the Tomiko Terrane is an area of relatively low peak pressures (*see* Figure 19.13b), and the Central Metasedimentary Belt Boundary Zone closely parallels the 6 to 8 kb and 8 to 10 kb boundary (*see* Figure 19.13b). Grant (1987, 1989) obtained P-T estimates that were on average lower than those of Anovitz and Essene (1990) and Anovitz (1987) (*see* Table 19.2), however the differences lie within the error range of P-T work.

### Distribution of Aluminosilicate Minerals

Distribution of aluminosilicate minerals is shown in Figure 19.13c. In the GFTZ, sillimanite is only abundant in the Beaverstone Domain (mainly in an area underlain by Killarney equivalent rocks; *see* Figure 19.13c). The kyanite-sillimanite boundary in the Central Gneiss Belt is not well-defined (*see also* Anovitz and Essene 1990) nor does it correlate directly with terrane boundaries (although in the west, it coincides roughly with the Parry Sound–Britt boundary; *see* Figure 19.13c). Davidson (1991) notes local-scale complications in the distribution of kyanite and sillimanite, most notably that according to the data of Anovitz and Essene (1990), all of the Central Gneiss Belt should be in the kyanite zone, and the isograd should lie close to the boundary with the Central Metasedimentary Belt. Alternatively, Davidson (1991) suggests that if sillimanite is the stable phase in the southern Central Gneiss Belt, then the P-T estimates may be in error.

### Distribution of Metamorphic Orthopyroxene

Distribution of orthopyroxene associated with granulite-facies rocks of "Grenvillian" age is shown on Figure 19.13d. Metamorphic orthopyroxene is concentrated in the Parry Sound and Frontenac terranes, with patches scattered through the Algonquin Terrane. Based on simple evaluation of mineral assemblages, the Central Gneiss Belt in Ontario contains distinct areas of granulite facies in an otherwise upper amphibolite-facies terrane (Davidson and Morgan 1981), with only small areas of middle amphibolite facies (e.g., stable muscovite + quartz). The presence of orthopyroxene, commonly with clinopyroxene and garnet in the more mafic gneisses and coupled with green to brown colouration of feldspars in leucocratic rocks, allows field delineation of relatively continuous tracts of granulite-facies assemblages in the Algonquin Terrane and the Parry Sound Domain (*see* Figure 19.13d).

Smaller patches of granulite-facies rocks occur in parts of the Britt, Go Home and Rosseau domains. All of these areas, and in particular the southeast part of the Parry Sound Domain, exhibit retrogression at local and regional scales (Davidson 1991). Some of the Grenville-age ductile deformation took place under granulite-facies conditions, causing recrystallization of older granulite assemblage minerals without retrogression (White and Mawer 1986; van Breemen et al. 1986). The patchy distribution of granulite-facies rocks could be explained if most of the rocks now exposed in the Central Gneiss Belt were once at granulite grade, having been downgraded in the process of ductile flow and migmatization during the northwest-directed thrust displacement believed to be the predominant manifestation of Grenvillian orogeny in this region (Davidson 1991). In addition, younger dike intrusion and extensional shear zones have also retrograded granulite-facies rocks on a local scale (Davidson et al. 1984; Davidson 1991).

Older periods of granulite metamorphism have been reported from various parts of the Algonquin Terrane (e.g., Britt Domain, Culshaw et al. 1991, 1992), but for the most part, these have been overprinted, and largely obliterated, by the Grenvillian event. In addition, granulite-facies rocks of Archean age may be present in the northeastern sector of the Grenville Front Tectonic Zone; Archean granulites occur south of the Grenville Front near Val d'Or, Quebec (Indares and Martingole 1989, 1990a). Indares and Martingole (1984) note that peak metamorphic conditions throughout the Mont-Laurier Terrane in Quebec attained granulite facies, although much of the terrane was subsequently re-equilibrated to upper amphibolite facies. A similar P-T history is likely in the eastern Bancroft Terrane in Ontario.

### Rare Metamorphic Minerals and Eclogitic Rocks

Figure 19.13e shows the distribution of rare metamorphic minerals from the Grenville Province. Of greatest interest in the Central Gneiss belt are the occurrences of sapphirine and pseudoeclogite assemblages (Davidson et al. 1982; Davidson 1990a, 1991; Grant 1987, 1989). Only 2 sapphirine occurrences are known, both in the Algonquin Terrane, within metabasic lenses (Davidson et al. 1982; Grant 1989).

Davidson (1990a, 1991) has drawn attention to eclogitic rocks within the Algonquin Terrane of the Central Gneiss Belt (*see* Figure 19.13e) originally considered pseudoeclogites (Davidson et al. 1982; Culshaw et al. 1983) because they lacked omphacite. Davidson (1990a, 1991) provides evidence that they were indeed eclogites (omphacite-garnet-kyanite rocks) prior to "retrogression." These rocks are restricted to ductile shear zones within the Algonquin Terrane and occur mainly in the high pressure dome shown by Anovitz and Essene (1990) (10 to 12 MPa area in Figure 19.13b).

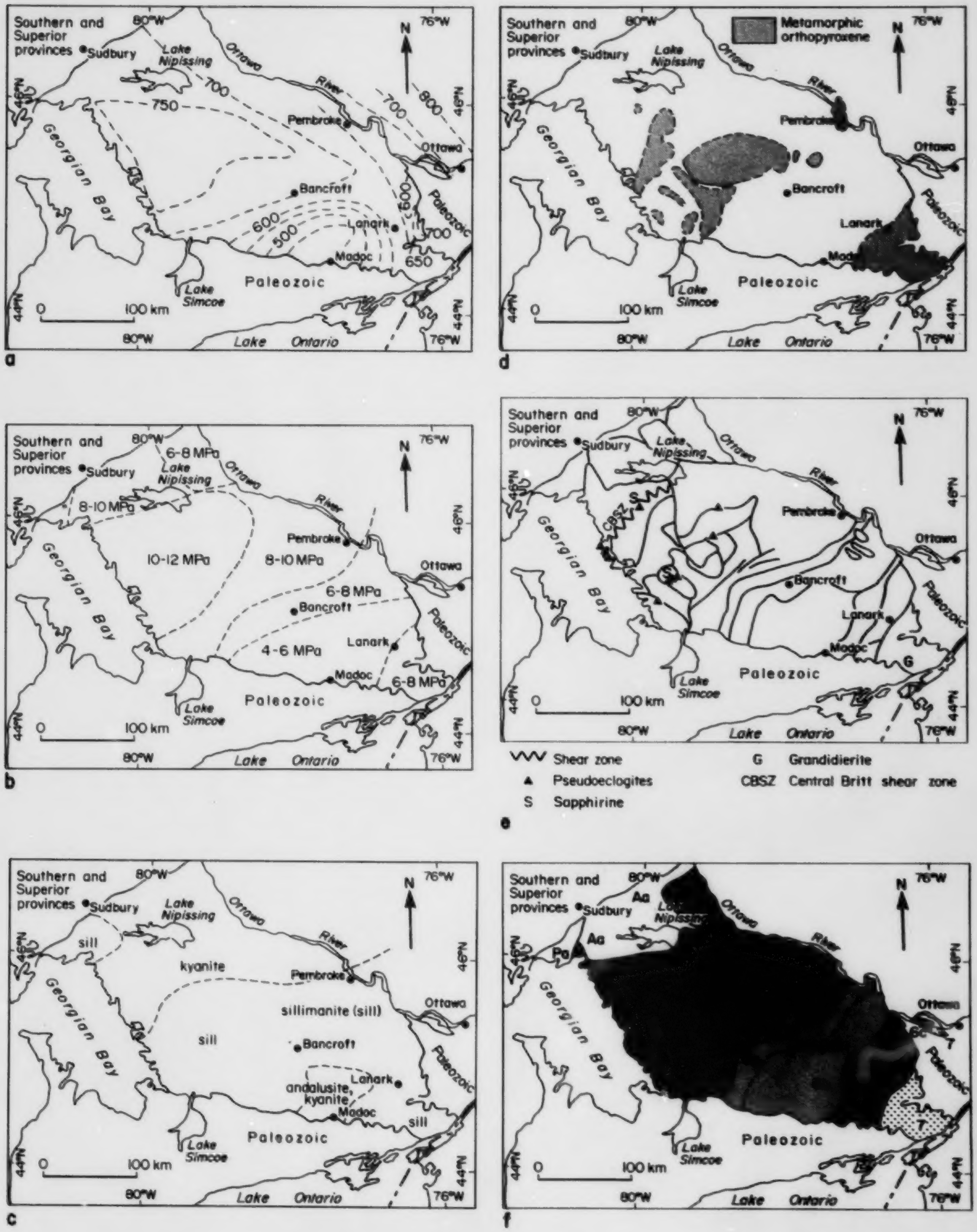

a
b
c
d
e
f
Southern and Superior provinces
Sudbury
Lake Nipissing
Ottawa River
Pembroke
Ottawa
Bancroft
Lanark
Madoc
Georgian Bay
Lake Simcoe
Paleozoic
Lake Ontario
80°W
76°W
46°N
44°N
N
0
100 km
750
700
800
600
500
650
6-8 MPa
8-10 MPa
10-12 MPa
4-6 MPa
sill
kyanite
sillimanite (sill)
andalusite, kyanite
Metamorphic orthopyroxene
CBSZ
S
G
Shear zone
Pseudoeclogites
Sapphirine
Grandidierite
CBSZ Central Britt shear zone
Aa
Pa
T

Figure 19.13. Regional metamorphic patterns across the Grenville Province in Ontario. a) Distribution of peak "Grenvillian" (1050 to 1200 Ma) metamorphic temperatures (°C). *Modified from* Anovitz and Essene (1990). b) Distribution of peak "Grenvillian" (1050 to 1200 Ma) metamorphic pressures. *Modified from* Anovitz and Essene (1990). c) Distribution of primary aluminosilicate minerals. *Modified from* Anovitz and Essene (1990). d) Distribution of metamorphic orthopyroxene ("Grenvillian"). *Modified from* Schau et al. (1986). e) Distribution of sapphirine, eclogitic rocks and other "exotic" metamorphic minerals. Terrane and domain boundaries as in Figure 19.2. *Modified from* Davidson et al. (1982), Culshaw et al. (1983) and Davidson (1990a). f) Stacking order of major lithotectonic "decks" (1 is on the bottom). Aa—Archean and Paleoproterozoic parautochthon; Pa—Paleo- to Mesoproterozoic parautochthon. *Modified from* Davidson (1984b), Schau et al. (1986) and Easton (1990a).

Grant (1989) estimated P-T conditions of about 14 to 16 kb and temperatures of 775° to 875°C from one occurrence. She suggested tectonic emplacement for the sapphirine-bearing and related eclogitic rocks involving rapid transport from depths of about 55 km in thickened crust to the higher crustal levels (25 to 30 km) represented by the preserved Central Gneiss Belt. Davidson (1990a, 1991), on the other hand, suggests that the surrounding quartzofeldspathic rocks were subjected to the same high-pressure conditions, and that they re-equilibrated to slightly lower pressure conditions as a result of continuous ductile strain during regional thrusting.

Carmichael et al. (1987) report an occurrence of grandidierite, an aluminous-borosilicate formed by anatectic reactions that consume tourmaline. If pressures had been higher in the Frontenac Terrane, kornerupine would have formed instead of grandidierite.

Figure 19.13f shows the various "decks" present within the Grenville Province which reflect the crustal-stacking order inferred to have occurred during northwestward-directed thrusting during the Grenville Orogeny (Davidson 1984b; Schau et al. 1986). The distribution of metamorphic grade and characteristic minerals is related to these decks (compare Figure 19.13f and Figure 19.13d), as crustal thickening will influence the P-T conditions of regional metamorphism.

## Areas of Low Grade Metamorphism

The area of lowest grade metamorphism in the Grenville Province, generally upper greenschist facies, in the Madoc-Bancroft area, has long been termed the "Hastings Metamorphic Low" (*see* Figures 19.13a and 19.13b). A less well known area of upper greenschist-facies metamorphism occurs in the Sharbot Lake Terrane north of Lanark (Figure 19.13a). Details of metamorphism in the Central Metasedimentary Belt will be discussed later; detailed metamorphic studies are listed in Table 19.1.

## Timing of Metamorphic Events

Metamorphism was diachronous across the major subdivisions of the Grenville Province as shown in Figure 19.14, and metamorphic events outlined for one terrane within the Grenville Province may have no counterparts in nearby parts of the province. It is also apparent from Figure 19.14 that it is difficult to define the nature and timing of the "Grenville metamorphic event" or the Elzevirian and Ottawan orogenies.

Lumbers et al. (1990) consider the main "Grenvillian" (Elzevirian) metamorphism in the Central Metasedimentary Belt as occurring between 1240 to 1180 Ma, and that the commonly cited *circa* 1100 to 1000 million-year-old metamorphic age reflects several unrelated local metamorphic and deformation events. Similar problems exist in interpreting the timing of metamorphism in the Central

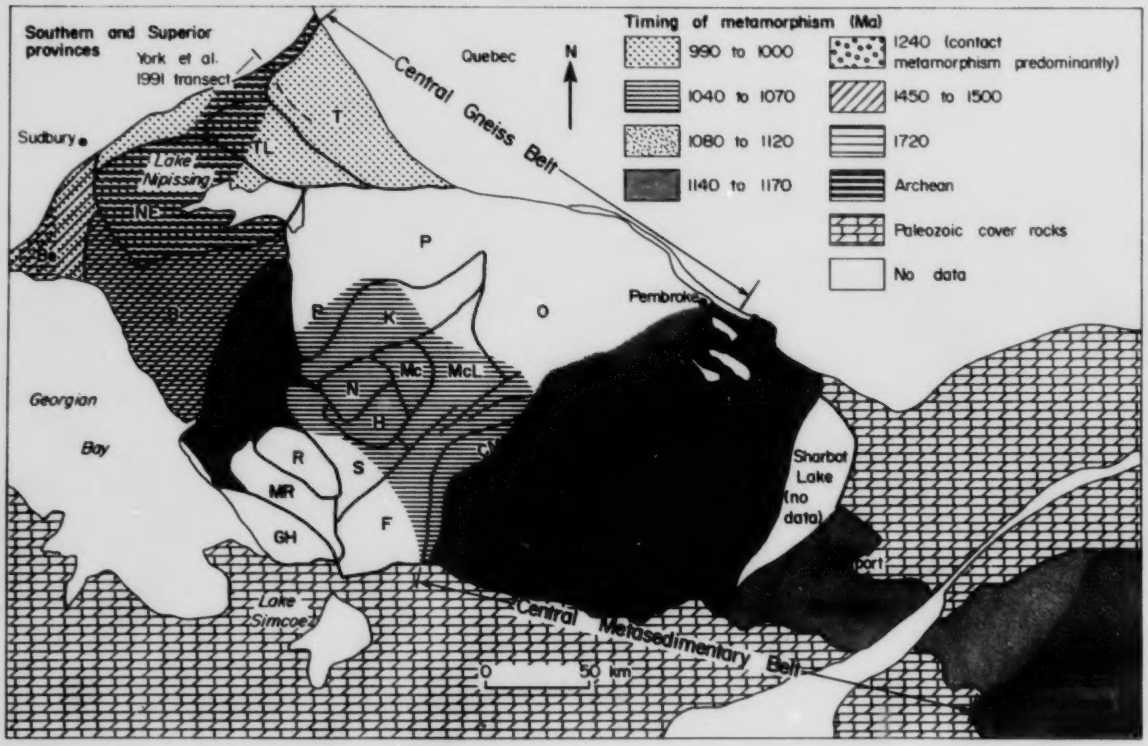

Southern and Superior provinces
York et al. 1991 transect
Quebec
N
Central Gneiss Belt
T
TL
Sudbury
Lake Nipissing
NE
B
P
K
O
Pembroke
Mc
McL
N
H
C
Georgian Bay
R
S
MR
GH
F
Sharbot Lake (no data)
port
Lake Simcoe
Central Metasedimentary Belt
0 50 km
Timing of metamorphism (Ma)
990 to 1000
1040 to 1070
1080 to 1120
1140 to 1170
1240 (contact metamorphism predominantly)
1450 to 1500
1720
Archean
Paleozoic cover rocks
No data

**Figure 19.14.** Time of metamorphic episodes across the Grenville Province based on U-Pb zircon, titanite and monazite ages. Terrane and domain boundaries as listed in Figure 19.2.

Gneiss Belt. Again, the main metamorphic episode may have taken place in the 1180 to 1160 million year period, with slow cooling resulting in the ubiquitous *circa* 1050 to 1000 million year ages throughout the region. Many granite pegmatites are postmetamorphic and have ages in the 1050 to 1000 million year range and therefore are not good indicators of the timing of regional metamorphism. For example, in the Central Metasedimentary Belt, a peak metamorphic episode may have affected all terranes within the Central Metasedimentary Belt at *circa* 1230 to 1180 Ma. A second metamorphic event, roughly coincident with the Ottawan Orogeny of *circa* 1140 to 1080 million year age, is well documented in the Mazinaw Terrane, where it affects both the Grenville Supergroup and the *circa* 1157 million-year-old (Kinsman and Parrish 1990) Flinton Group. The extent of this second metamorphic event across the Central Metasedimentary Belt is not well-constrained.

Subsequent cooling varied between the terranes, with the Frontenac Terrane passing through the argon blocking temperature in hornblende at *circa* 1120 Ma, whereas rocks in the Mazinaw and Bancroft terranes did not do so until *circa* 1000 Ma. In addition, a *circa* 1070 to 1030 alkaline metasomatizing and granite pegmatite emplacement event may have prevented early cooling in the Bancroft Terrane.

## Granulite-Facies Metamorphism

Two types of granulite terranes are present in the Grenville Province in Ontario, one type within the Central Gneiss Belt; the other in the Frontenac Terrane of the Central Metasedimentary Belt.

### CENTRAL METASEDIMENTARY BELT GRANULITES

This granulite terrane is similar to the metasedimentary granulite type of Percival (1989), in having almost congruent plutonism (1175 to 1165 Ma, Marcantonio et al. 1990, van Breemen and Davidson 1988b) and metamorphism (4 to 6 kb, 700° to 800°C at *circa* 1170 Ma, Mezger et al. 1990), sedimentation to within possibly 20 million years prior to plutonism and rapid uplift and cooling (Flinton Group contains detritus of Frontenac plutons at 1157±3 Ma, Kinsman and Parrish 1990). As shown in Figure 19.15, metamorphic conditions peak near Westport, and decrease to the north and south (Wynne-Edwards 1967a, 1967b). Dissenting views are expressed by Carmichael et al. (1978, 1987) and Lonker (1980).

Easton (1988b) noted that metamorphic grade in the Sharbot Lake Terrane increased with proximity to the Frontenac Terrane. This increase suggests that the Frontenac metamorphic event affected the Sharbot Lake Terrane, and that the terrane boundary formed prior to peak metamorphism in the Frontenac Terrane.

### CENTRAL GNEISS BELT GRANULITES

Within the Central Gneiss Belt, granulite-facies metamorphism occurred at several times. Within the Britt Domain, *circa* 1450 million-year-old granulite-facies metamorphism has been overprinted by "Grenvillian" amphibolite-facies metamorphism (Culshaw et al. 1991, 1992). Within the Parry Sound Terrane, granulite-facies metamorphism has been dated at 1161±3 Ma by van Breemen et al. (1986). Within the Algonquin Terrane, granulite-facies metamorphism took place at $1030^{+50}_{-20}$ Ma (van Breemen et al. 1986). In the Parry Sound and Algonquin terranes, plutons similar in age to the metamorphic peak are unknown, and the bulk of magmatism in both the Parry Sound and Algonquin terranes occurred some 250 to 350 million years prior to "Grenvillian" peak metamorphic conditions.

Emslie and Hunt (1990) argued that "Grenvillian" (1100 to 1030 Ma) regional high-grade metamorphism was caused, with an appropriate time lag, by crustal heating through copious emplacement of plutonic complexes of the anorthosite-mangerite-charnockite-granite suite (AMCG suite) between 1160 and 1120 Ma. They also report AMCG magmatism at 1360 Ma and 1640 Ma. A peculiarity of the Central Gneiss Belt in Ontario is its lack of Grenville-age plutons of the AMCG suite (or of any other suite), although AMCG rocks are abundant in parts of the Grenville Province in Quebec. With the exception of the deformed and metamorphosed Mulock Granite northeast of North Bay, Central Gneiss Belt felsic plutons are all older than 1340 Ma. However, as outlined in Figure 19.14, regional metamorphism in the Central Gneiss Belt was indeed "Grenvillian." Apparently, the only "Grenvillian" igneous rocks (excluding pegmatite) are abundant but small, scattered bodies of 1170 million-year-old coronitic olivine gabbro (Davidson 1991; Davidson and van Breemen 1988; van Breemen and Davidson 1990). Although these bodies could not themselves have significantly affected the total heat budget, it could be argued that they originated from much larger masses of mafic magma injected into the lower crust; however, there is no geophysical evidence to support this possibility. As Davidson (1991) notes:

> ...it seems more appropriate to turn to mechanisms of heat introduction and redistribution during the process of crustal thickening by thrust-stacking and overriding of hot slices derived from deeper levels. Gravitational compaction of ductile crust thickened in such a way would seem to be a feasible means of deriving the narrow, continuous units of highly attenuated rocks (straight gneisses) that are so prevalent in parts of the Central Gneiss Belt. ... In summary, much work on metamorphism in the Central Gneiss Belt, preferably coupled with structural and geochronologic studies, is still needed before details of P-T-t paths for different domains can be woven into an overall tectonic tapestry.

# Argon Isotope Studies and the Cooling History of the Grenville Province in Ontario

## Cooling Histories

Argon isotope ages provide constraints on postmetamorphic cooling, uplift and erosion rates in orogenic belts. Several such studies (Berger and York 1981a, 1981b; Baski 1982; Lopez-Martinez and York 1983; York 1984; Hanes et al. 1988; Anderson, S.L. 1988; Cosca 1989) across the Grenville Province help constrain tectonic models of the

**Table 19.1.** Metamorphic studies, Grenville Province, Ontario.

| | |
|---|---|
| **Regional Studies** | |
| Chesworth (1972) | Regional examination of distribution of aluminosilicate minerals in the Grenville Province in Ontario. |
| Bourne (1978) | Regional summary of Grenville metamorphism, includes brief section on Ontario. |
| Fraser et al. (1978) | Metamorphic map of Canada. |
| Moecher (1988) | Scapolite phase equilibria in granulite gneisses, Parry Sound region. |
| Anovitz and Essene (1990) | Regional survey of P-T conditions across the Grenville Province, includes detailed geothermometry and geobarometry and P-T paths. *Modified from* Anovitz (1987). |
| **Grenville Front Tectonic Zone** | |
| Pearson (1959) | Kyanite occurrences, south of the Grenville Front in the Sudbury area. |
| Kwak (1967, 1971) | P-T conditions of kyanite formation near Sudbury. |
| Dewitt (1976) | Sphalerite geobarometry in Grenville marbles. |
| LaTour (1979, 1981b) | P-T conditions in rocks southwest of the Grenville Front near Coniston. |
| O'Donnell (1986) | Metamorphism across the Grenville Front, Killarney area. |
| Bethune (1989) | Study of metamorphism of Sudbury Swarm diabase dikes in the Beaverstone Domain. |
| **Central Gneiss Belt (CGB)** | |
| Mummery (1972) | Study of coronite textures in the Parry Sound Domain. |
| Thompson (1983) | Metamorphic reactions, Parry Sound Domain. |
| Manojlovic (1987) | Retrogression of granulite-facies rocks in the Huntsville area. |
| Grant (1987, 1988) | Study of metagabbros in the Algonquin Terrane, includes geothermometry and geobarometry. |
| Grant (1989) | P-T conditions of sapphirine formation in metabasic rocks. |
| Davidson (1990a, 1991) | P-T conditions, distribution and significance of eclogitic rocks in the CGB. Davidson (1991) provides a summary of metamorphism in the CGB. |
| Tuccillo et al. (1990) | Metamorphic growth in garnets from the Britt Domain, discussion of P-T paths. |
| **Bancroft Terrane and Central Metasedimentary Belt Boundary Zone (CMBBZ)** | |
| Eisenbery (1954), Lal (1966), Lal and Moorhouse (1969) | Description of cordierite-gedrite gneisses in the Haliburton area. |
| Chesworth (1971) | P-T conditions in the Haliburton Highlands based on mapped mineral assemblages and petrography. |
| Sobol (1973) | Study of marbles in the Bancroft area. |
| Grieve and Gittins (1975) | Corona textures in the Hadlington Gabbro. |
| Cermignani (1979) | Metamorphic reactions in alkalic gneisses, Haliburton–Bancroft area. |
| Millar (1983) | Cordierite-gedrite gneisses in Bagot Township. |
| **Elzevir, Mazinaw and Sharbot Lake Terranes** | |
| Jennings (1969) | P-T conditions using mineral assemblages, Apsley–Bancroft area. |
| Carmichael (1967, 1970) | Mapping of isograds in the Whetstone Lake area. |
| Trzcienski (1971) | Study of metapelites in the Whetstone Lake area. |
| Thompson (1972) | Mapping of isograds in the Flinton Group, Northbrook–Ompah area. |
| Sampson (1972) | P-T conditions near Madoc. |
| Hutcheon (1972), Hutcheon and Moore (1973) | Isograds, P-T conditions in carbonates in the Marble Lake area. |
| Cermignani (1971), Cermignani and Anderson (1973) | Mineral assemblages in carbonates near Tweed. |
| Rivers (1976) | P-T conditions and metamorphic textures in the Northbrook–Ompah area. |
| Ewert (1977) | Mineral assemblages, isoreaction grads in carbonate rocks of the Sharbot Lake and Frontenac terranes. |

**Table 19.1.** Metamorphic studies, Grenville Province, Ontario.

| | |
|---|---|
| Carmichael et al. (1978) | Regional distribution of isograds in the Whetstone Lake area and in the Northbrook–Ompah area, field trip guide. |
| LeAnderson (1978), LeAnderson and Munoz (1987) | Metamorphism of carbonates, Limerick Township. |
| Mannard (1982) | Metamorphic assemblages, Coxvale area. |
| LeClair (1982), Thompson and LeClair (1987) | Chloritoid-hornblende schists, Whetstone Lake and Bishop Corners area, Elzevir and Mazinaw terranes. |
| Theriault (1985) | Carbonate metamorphism, Sharbot Lake Terrane. |
| Anderson, K.J. (1988) | P-T conditions using mineral assemblages in the Clyde Forks area. |
| Ford (1989) | Metamorphic petrology and metamorphic evolution of ultramafic rocks from the Flinton Creek area, Mazinaw Terrane. |
| **Frontenac Terrane** | |
| Hay (1962, 1965) | Study of garnet-cordierite gneisses in the Westport area. |
| Reinhardt (1965, 1968) | Study of mineral equilibria in pelitic gneisses in the region. |
| Reinhardt and Skippen (1970) | Petrochemistry of Frontenac granulites. |
| Allen (1976) | Silicate-carbonate equilibria in carbonates, Tudor Township. |
| Ewert (1977) | *See* listing above under Elzevir Terrane. |
| Carmichael et al. (1978) | Regional distribution of isograds in the Frontenac Terrane, field trip guide. |
| Carmichael et al. (1987) | Regional distribution of isograds in the Frontenac Terrane, synthesis of earlier metamorphic studies, field trip guide. |
| Lonker (1979, 1980) | Study of metamorphic conditions in pelitic gneisses in the region. |
| Schau et al. (1986) | Summary of data on Frontenac granulites, field trip guide. |

Grenville Province, as well as the later history of the region. Figure 19.16 summarizes these studies, and Table 19.3 summarizes the uplift and cooling rates derived from them.

Figure 19.16 shows a remarkable similarity in the cooling curves throughout the Grenville Province. This uniformity of cooling rates and paths over the entire Grenville implies stable patterns of geotherms throughout uplift, cooling and erosion (Cosca 1989). As shown in Table 19.3, cooling rates lie mainly in the range of 1° to 4°C/Ma in the Central Gneiss Belt and the Bancroft Terrane and 2° to 4°C/Ma for the remainder of the Central Metasedimentary Belt. There are only slight differences between various sectors of the Central Metasedimentary Belt (*see* Figure 19.16). In part, this uniformity may be a product of irregular data distribution, thereby masking subtle differences.

Uplift rates calculated from argon-argon (Cosca 1989) and potassium-argon studies (Anderson, S.L. 1988) are in the 0.07 to 0.14 cm/Ma range, or about 2 orders of magnitude less than uplift rates calculated from the Alps or the Himalayas (Cosca 1989; *see also* discussion by Cosca et al. 1990 and Anderson 1990). Cosca (1989) suggested this difference may be due to the fact that uplift rates in modern orogenic belts are calculated over shorter periods, and that the rates calculated from the Grenville may reflect time-integrated rates more typical of orogenic belts. It should be noted that in the above studies, it is assumed that peneplanation of the Grenville Province was not complete until the Early Cambrian. If peneplanation was completed earlier, uplift and erosion rates would be higher.

## Argon-Argon Studies, Regional Structures and Lithotectonic Domains

In his study of argon isotope effects across the Central Metasedimentary Belt, Cosca (1989) found significant differences between terranes within the Central Metasedimentary Belt (Figure 19.17). Figure 19.17 has been modified from the latter study to include additional hornblende ages and subdivision of the Elzevir Terrane into its component tectonic entities. The existing argon-argon data for the Central Metasedimentary Belt, although limited, correlates well with lithotectonic boundaries.

There is considerable variation, in some cases over 200 million years, in hornblende Ar-Ar ages across the Central Metasedimentary Belt (*see* Figure 19.17a). If the area were uniformly cooled, uplifted and eroded, then only significant differences in surface elevation would create any age variation, with the highest elevations preserving the oldest ages. Such variation in surface elevation across the Central Metasedimentary Belt is minimal. The simplest explanation is that tectonic movement, either during or postcooling, created the pattern shown in Figure 19.17b. All 3 major breaks in exposure depth are coincident with tectonic boundaries in the Central Metasedimentary Belt. Two of these, the Bancroft shear zone (Carlson et al. 1990) and the Robertson Lake mylonite zone (Easton 1987b, 1988a, 1988b, 1988d, 1988f), are

**Table 19.2.** Pressure-temperature estimates based on geothermometry and geobarometry, Grenville Province, Ontario.

| | T°C | P(kb) | Source |
|---|---|---|---|
| **Grenville Front Tectonic Zone** | no data | 8–10 | Anovitz and Essene (1990) |
| **Nipissing Terrane** | 650–750 | 8–8.5 | Anovitz and Essene (1990) |
| **Tomiko Terrane** | <700 | 6–8 | Anovitz and Essene (1990) |
| **Algonquin Terrane** | 700–800 | 8–10 | Culshaw et al. (1991) |
| Britt Domain | 700–750 | 9–10 | Corrigan (1990); Key Harbor Gneiss, |
| | 735 | 8.7 | Culshaw et al. (1991); Nadeau Island Gneiss, |
| | 700–750 | 9–10.5 | Anovitz and Essene (1990) |
| Go Home Domain | 710–720 | 8.8–10.3 | Grant (1987, 1989) |
| | no data | 9.5–10.5 | Anovitz and Essene (1990) |
| Kiosk Domain | 750–825 | 9.5–10.5 | Anovitz and Essene (1990) |
| Huntsville Domain | 700 | 9.2–9.9 | Grant (1987, 1989) |
| | no data | 8.5–10.5 | Anovitz and Essene (1990) |
| McClintock Domain | 670–715 | 7.9–8.8 | Grant (1987, 1989) |
| | no data | 8.5–10 | Anovitz and Essene (1990) |
| McCraney Domain | 700–800 | 10–10.5 | Anovitz and Essene (1990) |
| Novar Domain | 700–800 | 10.5 | Anovitz and Essene (1990) |
| Opeongo Domain | 710–720 | 8.2–8.8 | Grant (1987, 1989) |
| | 700–750 | 8.25–9.25 | Anovitz and Essene (1990) |
| Rosseau Domain | 700–800 | 8.5–10 | Anovitz and Essene (1990) |
| **Parry Sound Terrane** | | | |
| Parry Sound shear zone | 700–750 | 8 | Culshaw et al. (1991) |
| | 650–725 | 8 | Anovitz and Essene (1990) |
| Parry Sound Domain | 690–710 | 9.1–9.8 | Grant (1987, 1989) |
| | 750–800 | 10–11 | Anovitz and Essene (1990) |
| Moon River Domain | 700–800 | 9.5–11 | Anovitz and Essene (1990) |
| Seguin Domain | 700–800 | 11–12 | Anovitz and Essene (1990) |
| **Bancroft Terrane** | 550–625 | 5.5–6.5 | Anovitz and Essene (1990) |
| **Elzevir Terrane** | 375–425 | 3–4.5 | Anovitz and Essene (1990) |
| **Mazinaw Terrane** | 450–550 | 4–5 | Anovitz and Essene (1990) |
| **Sharbot Lake Terrane** | no data | no data | |
| **Frontenac Terrane** | 600–700 | 4.25–6 | Anovitz and Essene (1990) |

postmetamorphic, extensional shear zones along which considerable uplift has occurred. van der Pluijm and Carlson (1989) calculated 5 to 6 km of uplift along the Bancroft shear zone and suggested the shear zone was active between 1020 and 935 Ma. Similar uplift probably occurred along the Robertson Lake mylonite zone, as it juxtaposes upper amphibolite-facies rocks of the Mazinaw Terrane against greenschist-facies rocks of the Sharbot Lake Terrane (Easton 1987b, 1988a). The Mooroton shear zone (Easton and Ford 1990a, in prep.) also appears locally to have significant late vertical displacement.

York et al. (1991) describe a $^{40}$Ar/$^{39}$Ar study on single grains of biotite and hornblende along a 40 km traverse across the Grenville Front, along Highway 11, in the Temagami area. Minerals in the vicinity of the Grenville Front commonly contain excess argon, resulting in anomalously high integrated ages, which vary with distance from the Grenville Front. The overall age distribution along the front forms an asymmetric wavelike pattern with 2 age maxima, one on either side of the Grenville Front (Figure 19.18). The wave-like pattern of apparent ages in the Temagami area is remarkably similar to that seen in a potassium-argon study across the Grenville Front at Chibougamau, Quebec (Wanless et al. 1970; *see* Figure 19.18). Excess argon effects have been found along the Central Metasedimentary Belt Boundary Zone in the Minden area (Easton and Roddick 1988), and excess argon may be more common along ductile shear boundaries in the Grenville Province than previously recognized.

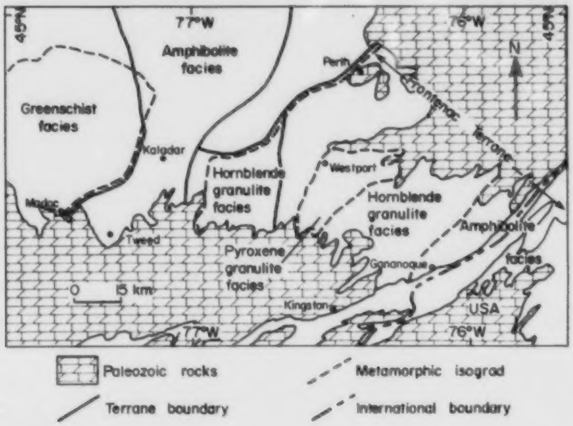


**Figure 19.15.** Distribution of granulite-facies rocks in the Frontenac Terrane, showing metamorphic isoreaction-grads and metamorphic culmination in the Westport area. *Modified from* Carmichael et al. (1987, 1978) and Wynne-Edwards (1967b).

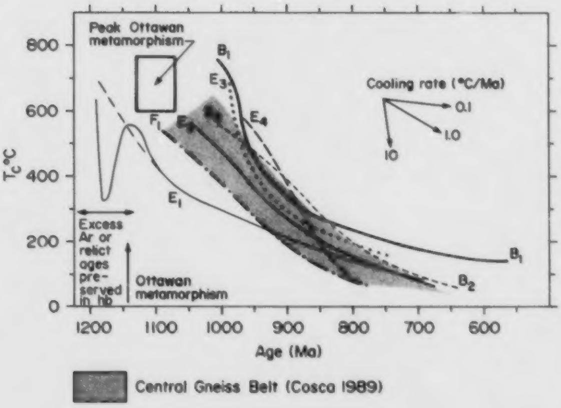


**Figure 19.16.** Summary diagram showing cooling age curves determined from argon-argon studies across the Grenville Province in Ontario. Stippled area shows range of Central Gneiss Belt determinations from Cosca (1989). B1—Bancroft Terrane curve *of* Berger and York (1981a, 1981b), York (1984); B2—Bancroft Terrane curve *of* Cosca (1989); E1—western Elzevir Terrane curve *of* Berger and York (1981a, 1981b), Lopez-Martinez and York (1983), York (1984); E2—western Elzevir Terrane curve *of* Cosca (1989); E3—central Elzevir Terrane curve *of* Hanes et al. (1988); E4—eastern Elzevir Terrane curve *of* Cosca (1989); F1—Frontenac Terrane curve *of* Cosca (1989).

**Table 19.3.** Uplift and cooling rates from the Grenville Province in Ontario.

| Uplift Rate (km/Ma) | Cooling Rate (°C/Ma) | Source and/or Comments |
|---|---|---|
| 0.07–0.14 | 2–4 | Cosca (1989) |
| 0.03–0.07 | no data | Cosca (1989), assumes peneplanation completed at 500 Ma. |
|  | 1.1–1.4 | Parry Sound area, Cosca et al.(1990), Anderson (1990) |
|  | 1.4–2.7 | Northern Algonquin Terrane, Cosca et al. (1990), Anderson (1990) |
| 0.02–0.24 | 1.1–2.7 | Anderson (1990) |
| 0.14–0.28 | no data | Himalayas, Cosca (1989) |
| 0.07–5.5 | no data | Himalayas, Cosca et al. (1990), Anderson (1990) |
| 0.3–2.2 | no data | Alps, Cosca et al. (1990), Anderson (1990) |

# Timing of "Grenvillian" Deformation

## Introduction

Figure 19.19 shows the distribution of major structural features across the Grenville Province. Domain- and terrane-bounding ductile shear zones, generally characterized by widespread development of straight and irregularly layered gneisses and porphyroclastic gneiss, represent recrystallized deep-crustal level mylonite zones (Davidson et al. 1982; Culshaw et al. 1983; Hanmer and Ciesielski 1984). Younger mylonite zones generally postdate the peak of the Grenville Orogeny (e.g., Robertson Lake mylonite zone, Easton 1987b, 1988a; Bancroft shear zone, van der Pluijm and Carlson 1989, Carlson et al. 1990) and are shown on Figure 19.19. These younger mylonite zones are commonly coincident, in the Central Metasedimentary Belt, with major stratigraphic and lithologic boundaries and may represent reactivated ductile shear zones. In the Central Metasedimentary Belt, a number of ductile deformation zones do not show the development of straight and porphyroclastic gneisses or subsequent recrystallization. These deformation zones are analogous to the deformation zones present in many Archean greenstone belts.

## Central Gneiss Belt

Figure 19.19 also shows U-Pb zircon ages along various shear zones across the Grenville Province, based on analysis of late-deformational pegmatites within the shear zones. Evidence of movement in the Grenville Front Tectonic Zone (Beaverstone Domain) post-1238 Ma is based on the age (Krogh et al. 1987) from the metamorphosed and deformed Sudbury Swarm diabase dikes (Davidson and Bethune 1988; Bethune and Davidson 1988; Bethune 1989). Zircon lower intercept ages and titanite ages from the area (Krogh 1989) give a minimum age of *circa* 995 million years on metamorphism in the GFTZ.

There is some suggestion of southeastward younging of deformation within the Central Gneiss Belt on the basis of the data shown in Figure 19.19; however, the data are incomplete and reactivation has occurred along these deformation zones (Nadeau and van Breemen 1990). Further, multiple deformation events are reported from the Central Metasedimentary Belt Boundary Zone in the Pembroke area at 1180 to 1140 Ma, 1060 Ma and as late as 1030 Ma (McEachern 1990; McEachern et al. 1990). McEachern et al. (1990) suggested that the *circa* 1060 million-year-old thrusting in the Haliburton area (van Breemen and Hanmer 1986) occurred out-of-sequence, and that *circa* 1180 to 1140 million-year-old thrusting occurred throughout the Central Metasedimentary Belt Boundary Zone, and possibly, the Central Gneiss Belt. At present, it appears certain that formation of the present stacking of domains and terranes in the Central Gneiss Belt was a complex process that was initiated as early as 1180 Ma, and persisted until *circa* 1030 Ma. Whether this was a continuous process, or a punctuated process (with peaks of activity at 1160, 1140, 1060 and 1040 to 1030 Ma), is still undetermined.

### Central Metasedimentary Belt

Timing of the development of domain and terrane boundaries in the Central Metasedimentary Belt is more poorly known, in part, because straight gneiss zones containing a variety of deformed pegmatites are rare, hampering efforts to obtain suitable material for dating deformation. In addition, it is difficult to precisely date deformation, unless deformed and undeformed pegmatites can be found in the same outcrop, as deformation effects may be localized.

The Frontenac Terrane records an 1170 to 1160 million-year-old metamorphic event that may not be present elsewhere in the Central Metasedimentary Belt. Note, however, that deformation may have been virtually complete prior to time recorded by the cooling of the metamorphic minerals dating this event. Further, the Flinton Group includes detritus from the Frontenac Terrane, suggesting that the Frontenac and Elzevir terranes became linked at that time. There is some overlap with the deformation peaks in the Central Gneiss Belt. The Frontenac Terrane accretion to the Elzevir Terrane at *circa* 1180 Ma is coincident with deformation in the Parry Sound shear zone.

Additional inferences can be drawn from the timing of metamorphism in various parts of the Central Metasedimentary Belt (*see* Figure 19.14). Both the Bancroft and Elzevir terranes record a metamorphic event at 1090 to 1070 Ma based on U-Pb titanite and monazite ages (e.g., Davis and Bartlett 1988; Mezger et al. 1990) suggesting some concurrency between the 2 terranes at that time. Metamorphism and deformation in the Central Metasedimentary Belt Boundary Zone and local deformation in the Central Gneiss Belt also occurred at 1090 to 1070 Ma (*see* Figure 19.19). The presence of late metamorphic to postmetamorphic syenite intrusions of *circa* 1085 Ma in all Central Metasedimentary Belt terranes (Corriveau 1989, 1990; Corriveau et al. 1990) seems to favour the peak of metamorphism being at 1090 to 1085 Ma, with cooling lasting for about 20 million years. As noted above in the argon-argon section, late-tectonic extensional movement is documented in the Central Metasedimentary Belt between 1000 and 900 Ma (van der Pluijm and Carlson 1989).

# CENTRAL GNEISS BELT

## Regional Patterns within the Central Gneiss Belt

### MAGMATISM

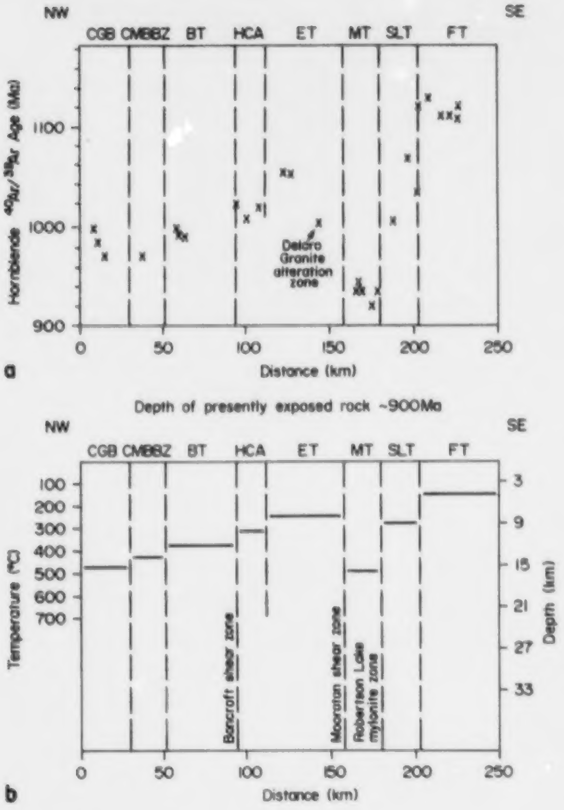
Five periods of magmatic activity have been recognized within the Central Gneiss Belt, however, only 2 volumetrically insignificant episodes can be considered "Grenvillian" in age (Figure 19.20). Little published geochemical data is available on these suites.

### 1740 to 1680 Million-Year-Old Activity

This event is not well-documented but corresponds roughly to the time of magmatic activity in the Killarney Magmatic Belt and the Mazatzal Belt of the midcontinent (Bickford et al. 1986; Easton 1986a). The 1740 million-year-old plutonism may be more widespread than previously appreciated (Easton 1992), as only recently has systematic mapping and dating of the older plutonic rocks of the Central Gneiss Belt, particularly in the Algonquin Terrane, begun (e.g., Culshaw et al. 1988, 1990, 1991). In the Nipissing Terrane, it is represented by the Sturgeon Lake and West Bay granites (Lumbers 1971a, 1975), which Lumbers (1978) and Lumbers and Vertolli (1991) consider to be roughly 1700 million years in age. Corrigan (1990) and Corrigan et al. (1990) have dated leucocratic gneisses in the Key Harbour area at *circa* 1684 Ma, and Nadeau and van Breemen (1990) report a $1714^{+123}_{-71}$ million-year-old U-Pb zircon age from a quartz monzonite in the Huntsville Domain. Dickin and McNutt (1989a) record 1800 million-year-old Nd-Sm model ages throughout the Algonquin Terrane, suggesting derivation of much of the crust in this region is from the mantle at *circa* 1800 to 1700 Ma. Mafic magmatism may also have occurred at this time, as Prevec (1992) reports an age of $1747^{+6}_{-5}$ million years from the Wanapitei gabbronorite intrusion located in the GFTZ southeast of Sudbury. The Wanapitei gabbronorite intrusion is a complex body consisting of a variety of phases, making interpretation of this new U-Pb age problematic.

### 1450 to 1420 Million-Year-Old Activity

This is the most widespread magmatic episode in the Central Gneiss Belt, corresponding to a period of widespread magmatism in the Eastern Granite–Rhyolite Province of the midcontinent (Bickford et al. 1986; Easton 1986a). Plutons of this age are found in all terranes of the Central Gneiss Belt (*see* Figures 19.20b and 19.2; *see* Map 2578, map case) and intrude crust having Nd-model ages of *circa* 1.8 Ga (Dickin and McNutt 1989a; Dickin et al. 1990). This also appears to have been a time of crust formation, as the Parry Sound, Seguin and Moon River domains all appear to be younger than *circa* 1450 Ma, as do the structural highest levels of the Britt and Go Home domains (*see also* Culshaw et al.



**Abbreviations**

| | | | |
|---|---|---|---|
| BT | Bancroft Terrane | FT | Frontenac Terrane |
| CGB | Central Gneiss Belt | HCA | Harvey-Cardiff Arch |
| CMBBZ | Central Metasedimentary Belt Boundary Zone | MT | Mazinaw Terrane |
| ET | Elzevir Terrane | SLT | Sharbot Lake Terrane |

**Figure 19.17.** Variation in Ar-Ar ages and inferences on the uplift history of the Central Metasedimentary Belt. **a)** Hornblende Ar-Ar ages plotted with respect to a cross section from Huntsville to Kingston across the Central Metasedimentary Belt, showing major tectonic divisions. *Modified from* Cosca (1989), includes additional data *from* Easton (1986a). **b)** Inferred burial depths at 900 Ma along a cross section from Huntsville to Kingston across the Central Metasedimentary Belt, showing major tectonic divisions. *Modified from* Cosca (1989), includes additional data *from* Easton (1986a) not found in Cosca (1989). Note significant differences across major late-extensional shear zones.

1990, 1991, 1992). Metamorphism is associated with this magmatic event in the Britt Domain (Culshaw et al. 1991), however, this early metamorphism is generally obliterated by the younger Grenvillian metamorphic overprint.

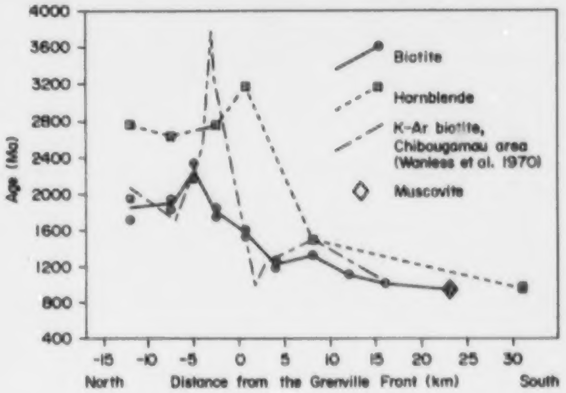


**Figure 19.18.** Variation in Ar-Ar ages from hornblende and biotite across the Grenville Front, south of Temagami, along Highway 11. Note excess argon peaks in biotite north and south of the Grenville Front showing an asymmetric wave-like pattern. Data obtained by Wanless et al (1970) on biotite in the Chibougamau area, Quebec is shown for comparison. Data *from* York et al. (1991).

## 1350 to 1320 Million-Year-Old Activity

There is increasing evidence for a widespread period of tonalite-granodioritic magmatism of a *circa* 1350 to 1320 million year age, at least in the southern Central Gneiss Belt, and perhaps extending south into the Central Metasedimentary Belt, the Adirondacks (Chiarenzelli and McLelland 1990, 1991; Daly and McLelland 1990) and Grenvillian inliers in the Appalachians (Aleinikoff et al. 1990; Ratcliffe and Aleinikoff 1990; Ratcliffe et al. 1991). Plutons of this age include the $1346^{+69}_{-39}$ million-year-old Nobel Granodiorite (van Breemen et al. 1986; 1330±44 Ma [Rb-Sr], Connare and McNutt 1985) located in the Britt Domain adjacent to the Parry Sound shear zone, the 1350±50 million-year-old anorthosites of the Parry Sound shear zone (van Breemen et al. 1986), the $1344^{+93}_{-32}$ million-year-old Redstone Lake Tonalite Gneiss (van Breemen and Hanmer 1986) and the *circa* 1350 million-year-old Dysart Tonalite Gneiss (Lumbers et al. 1990). van Breemen and Davidson (1990) also report a U-Pb zircon age of $1375^{+13}_{-12}$ million years from a metamonzonite gneiss from the Opeongo Domain. Lumbers and Vertolli (1991) also suggest that rocks of this age are locally present in the North Bay area, primarily in association with the Powassan Monzonite.

## 1250 Million-Year-Old Activity

The Mulock Granite, in the Tomiko Terrane, is the only pluton of this age known from the Central Gneiss Belt. It has a U-Pb zircon age of $1244^{+4}_{-3}$ Ma (Lumbers et al. 1991). This age corresponds to a period of widespread granite plutonism in the Elzevir Terrane of the Central Metasedimentary Belt. Basic magmatism associated with the Sudbury Diabase Dike Swarm is of similar age (*circa* 1238 Ma), but the relationship between these 2 magmatic events is unknown. As noted below, nepheline syenite suite rocks in the Bigwood Township area, and associated with the Eau Claire meta-anorthosite in Calvin Township (Lumbers 1976c), may represent magmatism of this age contemporaneous with nepheline syenite magmatism in the Central Metasedimentary Belt dated at *circa* 1290 to 1250 Ma (Lumbers et al. 1990). Prevec (1992) reports a U-Pb zircon age from the St. Charles meta-anorthosite of 1222±2 million years, however, his samples come from the deformed margin of the body, and may not accurately reflect the emplacement age of the body.

## 1170 Million-Year-Old Activity

Davidson (1990a, 1991), Davidson et al. (1982, 1985), Davidson and Grant (1986), van Breemen and Davidson (1990), Grant (1985, 1987) and Needham (1987) have all described a suite of basic gabbroic to locally ultramafic bodies, characterized by coronitic textures, throughout the Algonquin Terrane. These bodies are particularly abundant near domain boundaries (*see* Figure 19.20d). Baddeleyite from several widely scattered bodies yield ages of *circa* 1170 Ma (van Breemen and Davidson 1990), making this the only magmatic event of "Grenvillian-age" within the Central Gneiss Belt. This mafic magmatism is roughly coincident with plutonic activity in the Frontenac Terrane of the Central Metasedimentary Belt and may be related to accretion of the Frontenac Terrane to the Elzevir Terrane. Mafic magmatism of roughly similar age in North America is only known from the Superior Province; namely, the east-northeast-trending, 1140.6±2 million-year-old (Krogh et al. 1987) Abitibi dike swarm and the northeast-trending,

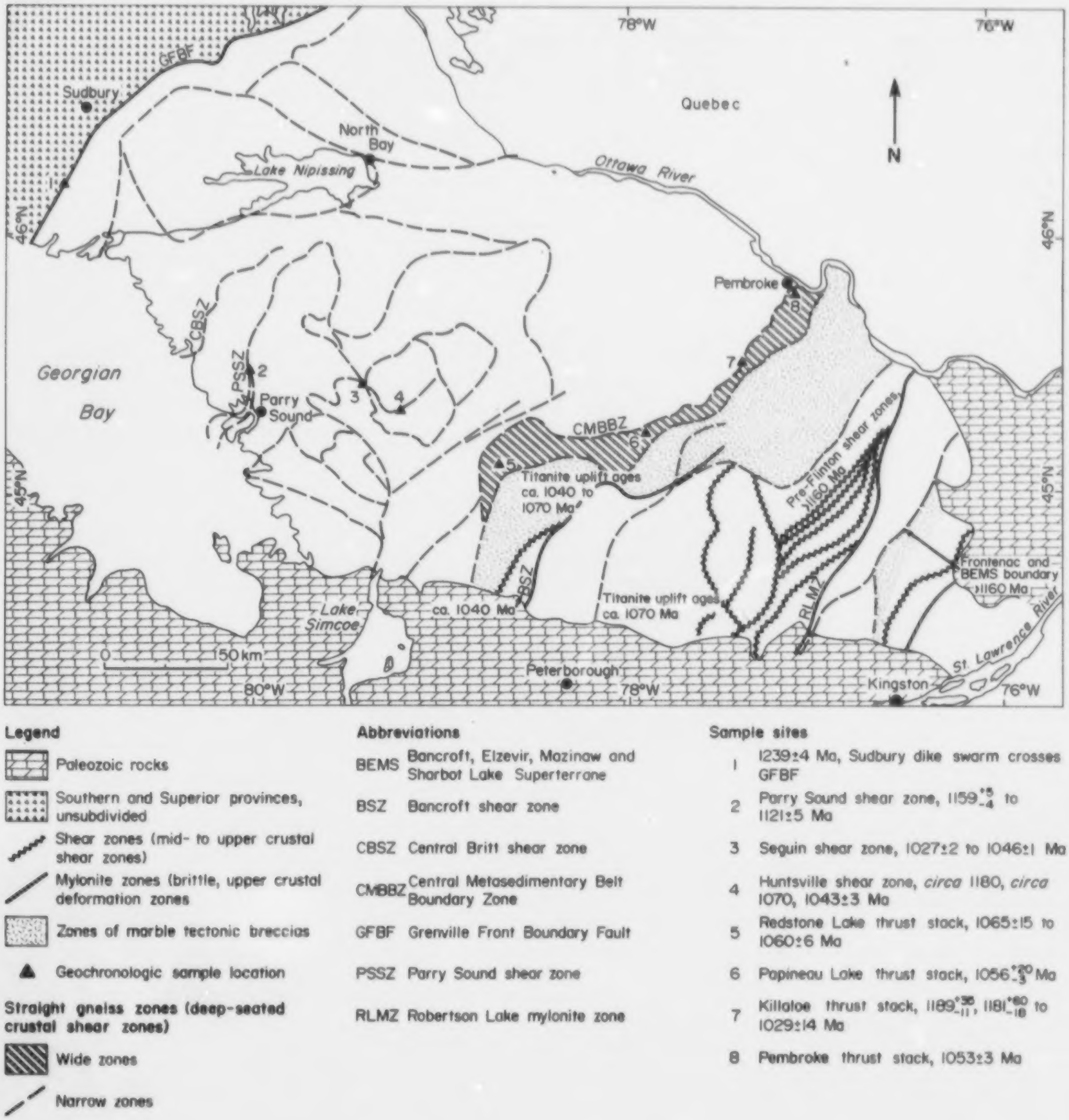

78°W
76°W
GFBF
Sudbury
North Bay
Quebec
N
Lake Nipissing
Ottawa River
1
46°N
46°N
Pembroke
8
7
CBSZ
PSSZ
2
Georgian Bay
Parry Sound
3
4
CMBBZ
6
5
Pre-Flinton shear zones >1160 Ma
Titanite uplift ages ca. 1040 to 1070 Ma
45°N
45°N
Frontenac and BEMS boundary >1160 Ma
BSZ
RLMZ
Titanite uplift ages ca. 1070 Ma
ca. 1040 Ma
Lake Simcoe
St. Lawrence River
0
50 km
Peterborough
80°W
78°W
Kingston
76°W
Legend
Paleozoic rocks
Southern and Superior provinces, unsubdivided
Shear zones (mid- to upper crustal shear zones)
Mylonite zones (brittle, upper crustal deformation zones
Zones of marble tectonic breccias
Geochronologic sample location
Straight gneiss zones (deep-seated crustal shear zones)
Wide zones
Narrow zones
Abbreviations
BEMS Bancroft, Elzevir, Mazinaw and Sharbot Lake Superterrane
BSZ Bancroft shear zone
CBSZ Central Britt shear zone
CMBBZ Central Metasedimentary Belt Boundary Zone
GFBF Grenville Front Boundary Fault
PSSZ Parry Sound shear zone
RLMZ Robertson Lake mylonite zone
Sample sites
1 1239±4 Ma, Sudbury dike swarm crosses GFBF
2 Parry Sound shear zone, $1159^{+5}_{-4}$ to 1121±5 Ma
3 Seguin shear zone, 1027±2 to 1046±1 Ma
4 Huntsville shear zone, *circa* 1180, *circa* 1070, 1043±3 Ma
5 Redstone Lake thrust stack, 1065±15 to 1060±6 Ma
6 Papineau Lake thrust stack, $1056^{+20}_{-3}$ Ma
7 Killaloe thrust stack, $1189^{+36}_{-11}$, $1181^{+60}_{-18}$ to 1029±14 Ma
8 Pembroke thrust stack, 1053±3 Ma

**Figure 19.19.** Timing of "Grenvillian" (1200 to 1050 Ma) deformation in the Grenville Province, Ontario. Terrane and domain boundaries as listed in Figure 19.2.

1135 million-year-old (Hansen 1975) Pigeon River dike swarm (*see* Osmani, this volume). In the Grenville Province in Quebec, most anorthosite massifs were emplaced at roughly 1160 million years, probably in an extensional regime (Gower et al. 1990; Higgins and Hervert 1992), indicating that by this time, collision associated with the Elzevirian Orogeny was complete.

## Granite Pegmatites

Pegmatite fields in the Central Gneiss Belt (Goad 1990) are illustrated in Figure 19.20d. These fields consist of granite and syenite pegmatites, which are locally uraniferous. The Conger Township and Blackstone Lake pegmatites south of Parry Sound have maximum $^{207}$Pb/$^{206}$Pb ages of *circa* 1017 Ma (Cumming et al. 1955; Davis et al. 1960; Tilton et al. 1960), similar to pegmatites in the Bancroft area (*see also* Table VII in Easton 1986a). Goad (1990) concluded that the pegmatites in the Grenville Province were likely the result of localized partial melting in areas of high metamorphic grade, and that their composition was not indicative of any significant rare-metal potential.

## Anorthosites

Although most reviews of the Grenville Province (e.g., Wynne-Edwards 1972; Baer 1976, 1981; Windley 1989; Emslie and Hunt 1990) emphasize anorthositic magmatism as a characteristic feature of the Grenville Province, anorthosites are poorly represented within the Central Gneiss Belt (Figure 19.21). No large massifs are present, and the anorthosites occur either as: 1) thin sills or dikes (e.g., St. Charles and Head Lake meta-anorthosites); 2) large bodies with tectonized margins in the western Parry Sound Terrane and in the Parry Sound shear zone; and 3) discontinuous lenses, pods and blocks within tectonic zones marking terrane boundaries. The largest concentration of anorthosite in the Central Gneiss Belt is spatially associated with the basal thrust of the Parry Sound Terrane (Parry Sound shear zone; *see* Figure 19.21), and includes the Whitestone, Parry Island, Arnstein and Raganooter gabbroic anorthosites.

Easton (1988e, 1990b) reviews the distribution and chemistry of meta-anorthosites in Ontario. Most anorthosites are either type 2 (massif-type anorthosites *of* Ashwal and Burke 1987) or type 3 (anorthosites associated with layered mafic intrusions *of* Ashwal and Burke 1987), or variants of these two types. Complex deformational histories and tectonic disruption, or lack of study, make it difficult to readily classify the meta-anorthosites found in the Central Gneiss Belt.

One of the larger meta-anorthosite complexes is the River Valley Anorthosite located adjacent to, and locally straddling, the Grenville Front northeast of Sudbury (*see* Figure 19.21; Lumbers 1973). Ashwal and Wooden (1989) conducted a detailed study of the body, obtaining a Pb-Pb whole-rock age of 2556±155 Ma, a Nd-Sm age of 2377±68 Ma and a Rb-Sr age of 2185±105 Ma. The latter 2 ages indicate slight resetting of the isotopic systems after emplacement. The body is dominated by coarse leuconorite and leucogabbro, with lesser anorthosite, gabbro and rare ultramafics, with plagioclase of $An_{60-70}$, all characteristic of type 3 anorthosites (Ashwal and Wooden 1989). In addition to the abundant igneous-textured rocks present in the main part of the body, a variety of deformed and metamorphosed equivalents of these rocks are present. Ashwal (Lunar and Planetary Institute, written communication, 1988) suggests that the River Valley Anorthosite may be related to other Paleoproterozoic intrusions in the region, such as the Shakespeare–Dunlop and the East Bull Lake gabbro-anorthosite intrusions, zircon dated at 2491±5 Ma and $2480^{+10}_{-5}$ Ma, respectively (Krogh et al. 1984). The Red Deer meta-anorthosite, located southwest of the River Valley Anorthosite, has not been studied in detail but may be either a sliver of the River Valley Anorthosite or a separate, but related body.

The Mercer and St. Charles meta-anorthosites (Lumbers 1975; Rousell 1978, 1981) are sills displaying massif anorthosite chemical affinity (Figure 19.23, Burwash area field in Figure 19.22) with preserved igneous textures. The Fall Lake, Pickerel Complex and Eau Claire meta-anorthosites are briefly described by Lumbers (1976c).

Four dominantly gabbroic anorthosite bodies, the Whitestone, Parry Island, Arnstein and Raganooter bodies occur in the Parry Sound Terrane (*see* Figure 19.21). The 1350±50 million-year-old Whitestone Gabbroic Anorthosite (van Breemen et al. 1986) has received considerable study, notably its petrology (Mason 1969), rare earth element (REE) geochemistry (Barker 1972), ore mineralogy (Kretschmar 1968) and industrial mineral potential (Marmont and Johnson 1987; Marmont 1988a). Mason (1969) concluded that the Whitestone Gabbroic Anorthosite was a massif-type anorthosite on the basis of its mineralogy and geochemistry.

The Arnstein and Raganooter meta-anorthosites lie along strike from the Whitestone Gabbroic Anorthosite (Marmont and Johnston 1987; Marmont 1988a). The Parry Island meta-anorthosite lies southwest of the Whitestone Gabbroic Anorthosite and has been studied by Marmont et al. (1988). All 4 masses occur on the sheared, western boundary of the Parry Sound Domain, and may represent slivers of a larger mass. In addition to these larger masses, some thin anorthosite units and inclusion trains of anorthositic rocks occur along the northern margin of the Moon River Domain (e.g., Moon River Bay meta-anorthosite, Figure 19.21) near the boundary with the Parry Sound Domain. These anorthosite units and inclusion trains lie at the same structural level as the larger anorthosite masses in the Parry Sound Domain and may also be sliced off the same body that gave rise to the Parry Sound Domain anorthosites (cf., Culshaw et al. 1989). Perhaps the Parry Sound Domain and Moon River Domain anorthosites are exotic (type 2 anorthosites), or perhaps they are genetically related to the mafic rocks which surround them (type 2, type 3, or possibly type 4 anorthosites *of* Ashwal and Burke 1987).

Tectonically disrupted meta-anorthosites elsewhere in the Central Gneiss Belt are shown on Figure 19.21. They are common along major tectonic boundaries, particularly the domain boundaries in the Central Gneiss Belt, and represent

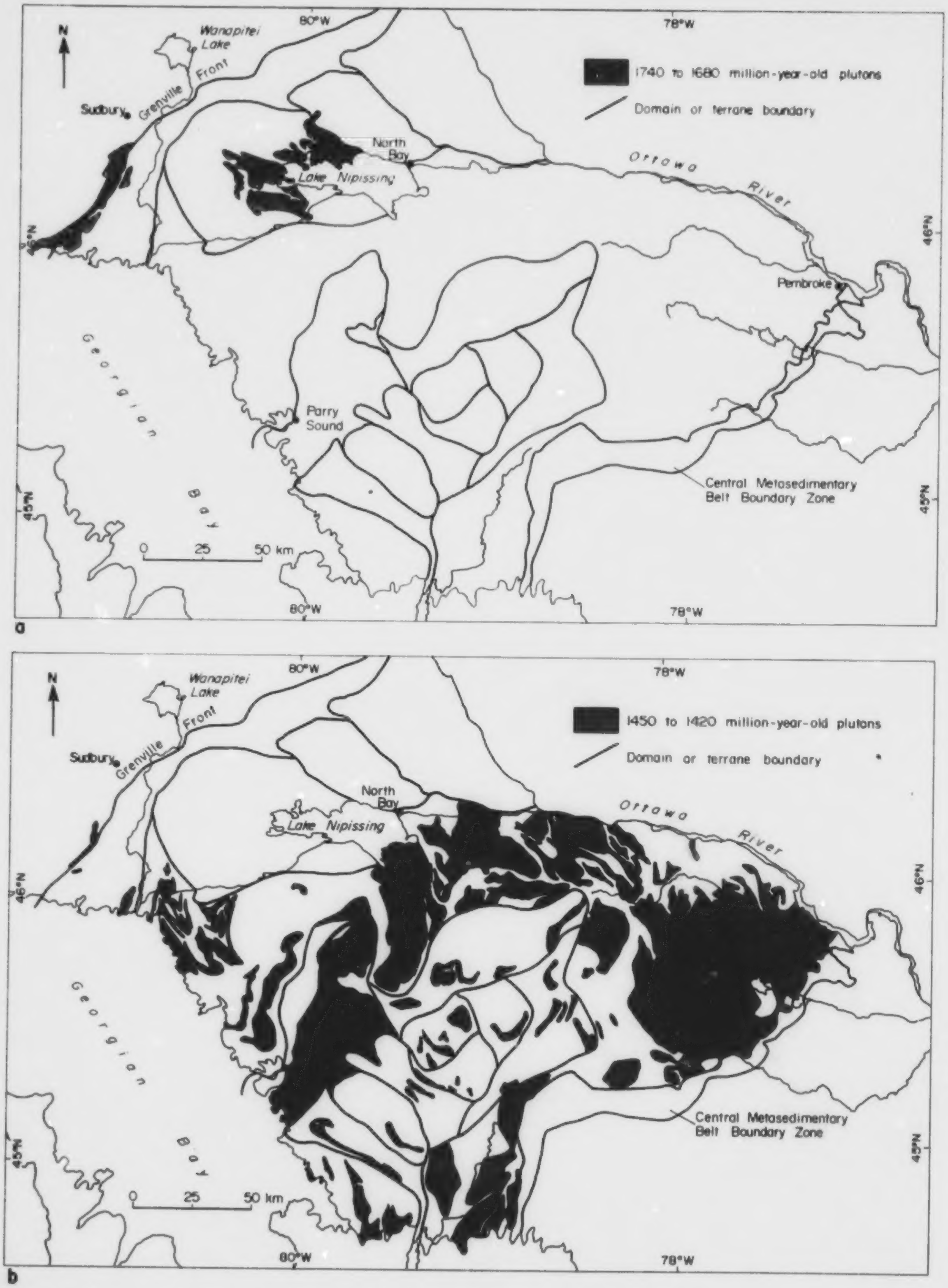

N
Wanapitei Lake
Sudbury
Grenville Front
80°W
78°W
1740 to 1680 million-year-old plutons
Domain or terrane boundary
North Bay
Lake Nipissing
Ottawa River
Pembroke
Parry Sound
Georgian Bay
Central Metasedimentary Belt Boundary Zone
46°N
45°N
0
25
50 km
a
1450 to 1420 million-year-old plutons
b

**Figure 19.20.** Distribution of plutons related to the main periods of magmatic activity in the Central Gneiss Belt. **a)** 1740 to 1680 million-year-old plutons; **b)** 1450 to 1420 million-year-old plutons; **c)** 1350 to 1320 million-year-old plutons; and **d)** 1250 and 1170 million-year-old plutons and location of major pegmatite fields. *See* Figure 19.2 for terrane and domain names.

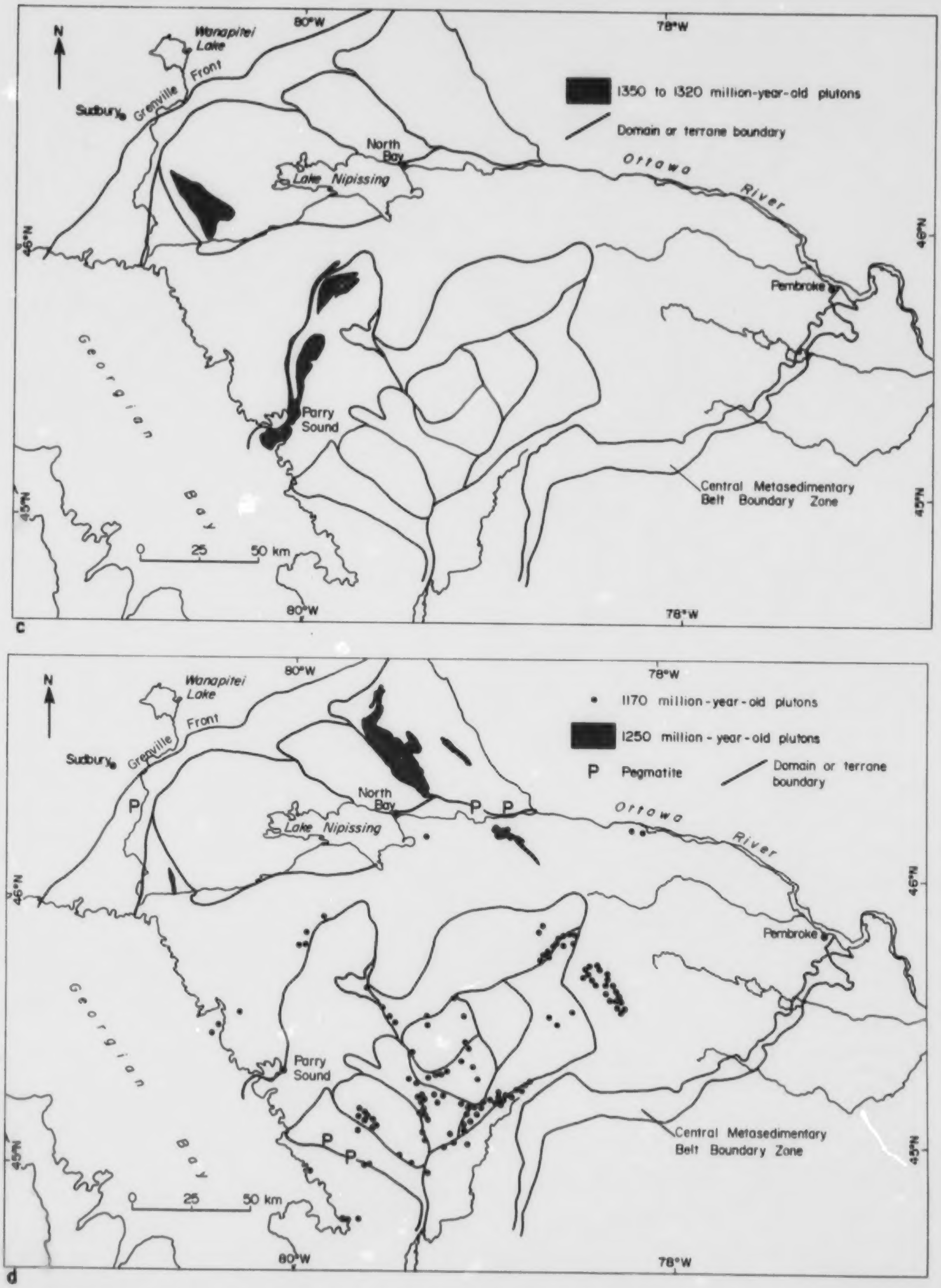

N
Wanapitei Lake
Sudbury
Grenville Front
80°W
78°W
1350 to 1320 million-year-old plutons
Domain or terrane boundary
North Bay
Lake Nipissing
Ottawa River
46°N
Pembroke
Georgian Bay
Parry Sound
Central Metasedimentary Belt Boundary Zone
45°N
0 25 50 km
c
N
Wanapitei Lake
Sudbury
Grenville Front
80°W
78°W
1170 million-year-old plutons
1250 million-year-old plutons
P Pegmatite
Domain or terrane boundary
North Bay
Lake Nipissing
Ottawa River
P
46°N
Pembroke
Georgian Bay
Parry Sound
Central Metasedimentary Belt Boundary Zone
45°N
0 25 50 km
d

**Figure 19.20.** Distribution of plutons related to the main periods of magmatic activity in the Central Gneiss Belt. **a)** 1740 to 1680 million-year-old plutons; **b)** 1450 to 1420 million-year-old plutons; **c)** 1350 to 1320 million-year-old plutons; and **d)** 1250 and 1170 million-year-old plutons and location of major pegmatite fields. *See* Figure 19.2 for terrane and domain names.

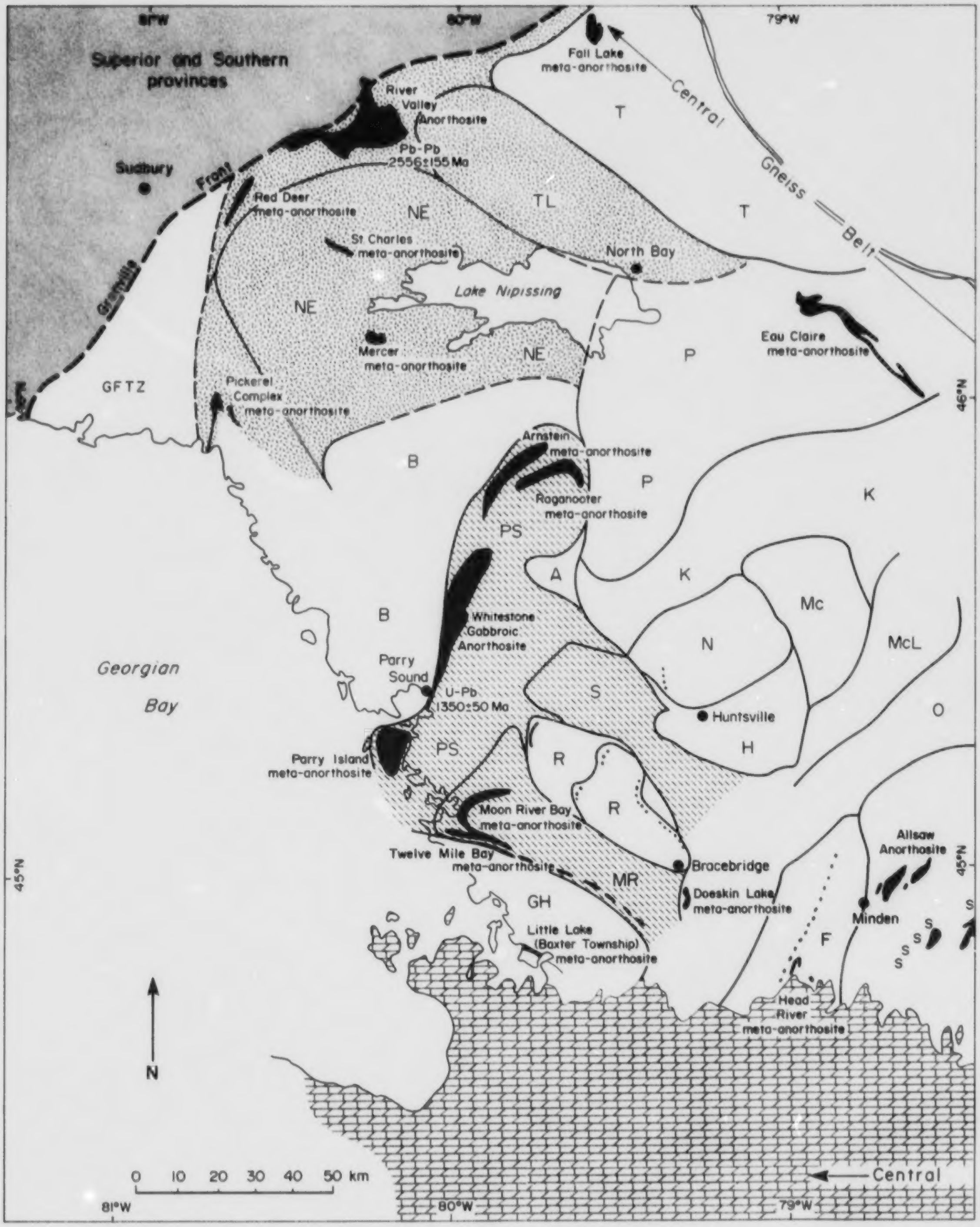

81°W
80°W
79°W
Superior and Southern provinces
Fall Lake meta-anorthosite
Central Gneiss Belt
River Valley Anorthosite
Pb-Pb 2556±155 Ma
T
TL
T
Sudbury
Front
Grenville
Red Deer meta-anorthosite
NE
St Charles meta-anorthosite
North Bay
Lake Nipissing
NE
Eau Claire meta-anorthosite
Mercer meta-anorthosite
NE
P
GFTZ
Pickerel Complex meta-anorthosite
46°N
B
Arnstein meta-anorthosite
P
K
Raganooter meta-anorthosite
PS
A
K
Mc
B
Whitestone Gabbroic Anorthosite
N
McL
Georgian Bay
Parry Sound
U-Pb 1350±50 Ma
S
Huntsville
O
H
Parry Island meta-anorthosite
PS
R
R
Moon River Bay meta-anorthosite
Allsaw Anorthosite
Twelve Mile Bay meta-anorthosite
Bracebridge
45°N
MR
GH
Doeskin Lake meta-anorthosite
Minden
Little Lake (Baxter Township) meta-anorthosite
F
S
S
S
S
Head River meta-anorthosite
N
0 10 20 30 40 50 km
Central
81°W
80°W
79°W

**Figure 19.21.** Distribution of metamorphosed anorthositic rocks and gabbroic intrusions within the Grenville Province, Ontario. Most anorthosites within the Central Metasedimentary Belt are in the Bancroft and Sharbot Lake terranes associated with gabbroic intrusions. Terrane and domain boundaries are listed in Figure 19.2. Crustal ages in the Central Gneiss Belt are based on Nd-Sm model-age data *of* Dickin and McNutt (1989a, 1989b, 1990) and existing geochronology information summarized in Easton (1986a). Southern limit of Archean and Paleoproterozoic crust in the Central Gneiss Belt is only approximate. Pb-Pb whole rock age on the River Valley Anorthosite *from* Ashwal and Wooden (1989); U-Pb zircon age on the Mill Lake Anorthosite (Whitestone Gabbroic Anorthosite) *from* van Breemen et al. (1986). The Tallan Lake Sill amphibolite has yielded a U-Pb zircon age of 1243 Ma (Heaman et al. 1986).

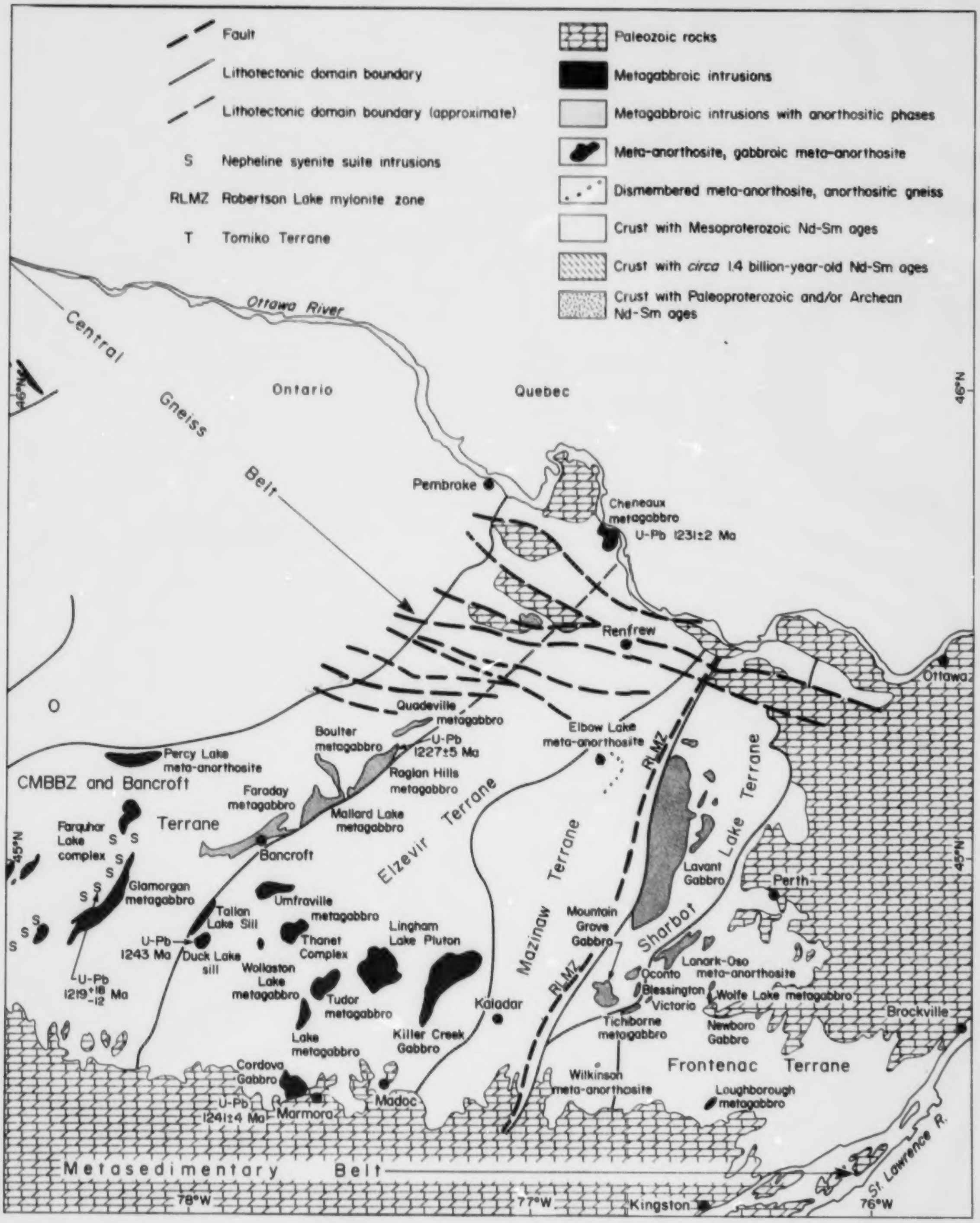

Fault
Lithotectonic domain boundary
Lithotectonic domain boundary (approximate)
S Nepheline syenite suite intrusions
RLMZ Robertson Lake mylonite zone
T Tomiko Terrane
Paleozoic rocks
Metagabbroic intrusions
Metagabbroic intrusions with anorthositic phases
Meta-anorthosite, gabbroic meta-anorthosite
Dismembered meta-anorthosite, anorthositic gneiss
Crust with Mesoproterozoic Nd-Sm ages
Crust with circa 1.4 billion-year-old Nd-Sm ages
Crust with Paleoproterozoic and/or Archean Nd-Sm ages
Ottawa River
Central Gneiss Belt
Ontario
Quebec
46°N
Pembroke
Cheneaux metagabbro
U-Pb 1231±2 Ma
Renfrew
Ottawa
O
Quadeville metagabbro
Boulter metagabbro
U-Pb 1227±5 Ma
Raglan Hills metagabbro
Percy Lake meta-anorthosite
CMBBZ and Bancroft Terrane
Faraday metagabbro
Mallard Lake metagabbro
Elbow Lake meta-anorthosite
RLMZ
Elzevir Terrane
Sharbot Lake Terrane
Lavant Gabbro
Perth
45°N
Farquhar Lake complex
Bancroft
Glamorgan metagabbro
Tallan Lake Sill
Umfraville metagabbro
U-Pb 1243 Ma
Duck Lake sill
Thanet Complex
Lingham Lake Pluton
Mazinaw Terrane
Mountain Grove Gabbro
Lanark-Oso meta-anorthosite
Oconto
Blessington
Victoria
Wolfe Lake metagabbro
U-Pb 1219 +18 -12 Ma
Wollaston Lake metagabbro
Tudor metagabbro
Kaladar
Tichborne metagabbro
Newboro Gabbro
Brockville
Lake metagabbro
Killer Creek Gabbro
Frontenac Terrane
Cordova Gabbro
U-Pb 1241±4 Ma
Marmora
Madoc
Wilkinson meta-anorthosite
Loughborough metagabbro
Metasedimentary Belt
78°W
77°W
Kingston
76°W
St. Lawrence R.

**Figure 19.21.** Distribution of metamorphosed anorthositic rocks and gabbroic intrusions within the Grenville Province, Ontario. Most anorthosites within the Central Metasedimentary Belt are in the Bancroft and Sharbot Lake terranes associated with gabbroic intrusions. Terrane and domain boundaries are listed in Figure 19.2. Crustal ages in the Central Gneiss Belt are based on Nd-Sm model-age data *of* Dickin and McNutt (1989a, 1989b, 1990) and existing geochronology information summarized in Easton (1986a). Southern limit of Archean and Paleoproterozoic crust in the Central Gneiss Belt is only approximate. Pb-Pb whole rock age on the River Valley Anorthosite *from* Ashwal and Wooden (1989); U-Pb zircon age on the Mill Lake Anorthosite (Whitestone Gabbroic Anorthosite) *from* van Breemen et al. (1986). The Tallan Lake Sill amphibolite has yielded a U-Pb zircon age of 1243 Ma (Heaman et al. 1986).

mylonitic, boudinaged and inclusion trains of meta-anorthosite, anorthosite gneiss and gabbroic anorthosite gneiss.

Van Kranendonk (1985, 1987, 1991) and Easton (1987e) described a series of thin meta-anorthosite bodies, (*see* Figure 19.21, Baxter Township meta-anorthosites and Head River meta-anorthosite) not clearly associated with any large-scale tectonic structures. They consist of thin (50 to 200 m wide), but extensive (up to 15 km long) units, possibly originally emplaced as sheets of anorthosite, gabbroic anorthosite and anorthositic gabbro. Locally, igneous layering is preserved in these bodies. In overall composition, they resemble massif-type anorthosites (Van Kranendonk 1987, 1991; *see* Figures 19.22 and 19.23). Van Kranendonk (1987, 1991) concluded that these units were intrusive, and that they were differentiates from a gabbroic anorthosite parent. A suite of metamorphosed monzodiorite and quartz diorite intrusions are associated with the Head River meta-anorthosite (*see* Figure 19.21), and these may have been comagmatic with the anorthositic magmatic event (Van Kranendonk 1987). These thin meta-anorthosite units may represent a subclass of massif-type anorthosites. No age determinations have been made on these thin meta-anorthosite units, but based on comparisons of the geologic history of the Fishog Domain with other better dated parts of the Central Gneiss Belt, an emplacement age of 1400 to 1300 Ma is likely (Easton 1987e).

## TECTONITE NOMENCLATURE AND PROTOLITH RECOGNITION

### Tectonite Nomenclature

Many rocks in the Central Gneiss Belt have been subjected to extreme ductile deformation and are best described as tectonites. Davidson et al. (1982, 1984), Culshaw et al. (1983) and Hanmer and Ciesielski (1984) introduced a number of field-based terms to describe these tectonic gneisses; some of which are defined below, as they are not present in the Glossary of Geology (Bates and Jackson 1987).

#### TRANSPOSED GNEISS

A transposed gneiss is continuous (over tens of metres), parallel-sided and has centimetre-thick layers of very fine-grained, leucocratic granitoid and amphibolite gneiss with layers of more intermediate composition (Figures 19.24 and 19.25). Transposed gneisses may locally grade into straight gneisses.

#### STRAIGHT GNEISS

A straight gneiss is a very fine-grained, well-layered gneiss whose layering is not as rectilinear as that in a transposed gneiss, but which contains few discordant features typical of irregularly layered gneisses (Figure 19.26). It may be intermixed with irregularly layered gneiss, as later deformation will cause local discordance in what was originally a straight or transposed gneiss (*see* Figure 19.26).

#### BLOCK GNEISS

Block gneisses generally consist of amphibolite blocks or layers in a less competent quartzofeldspathic gneiss (Figures 19.27 and 19.28).

#### GRANULAR GNEISS

Granular gneiss is a new term introduced herein, to describe medium-grained, protomylonitic gneisses with isolated, yet abundant, feldspar porphyroclasts. It lacks the fine-grained matrix characteristic of porphyroclastic gneisses, and feldspar is more abundant and more closely packed than in porphyroclastic gneisses (Figures 19.28 and 19.29).

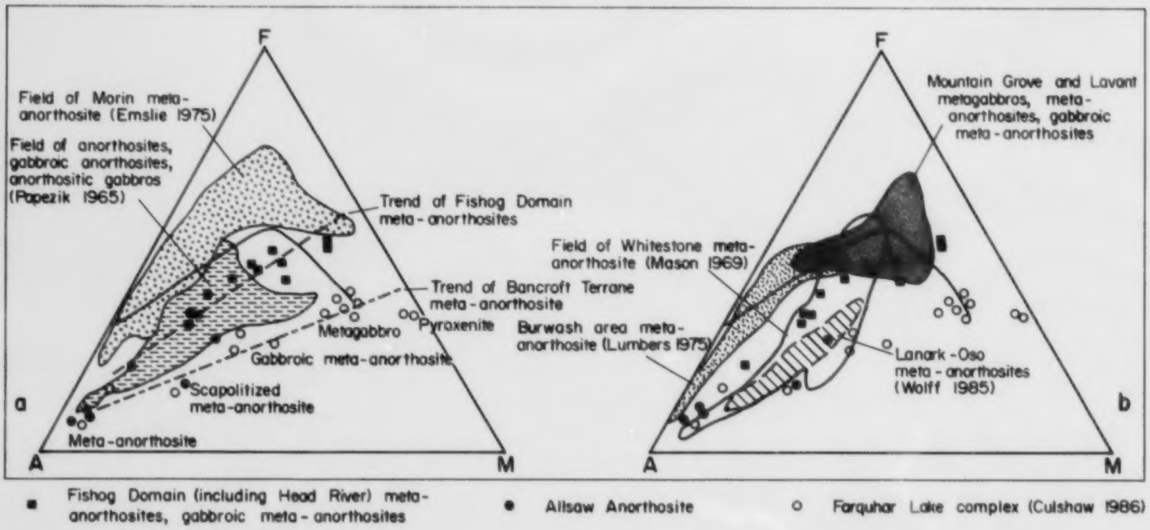


**Figure 19.22.** a) AFM plot of meta-anorthositic rocks from the Grenville Province, Ontario compared to other anorthositic units. Note distinct field and trend of low ferromagnesian meta-anorthosites of the Bancroft Terrane. b) AFM plot of anorthositic rocks from the Grenville Province, Ontario. The Lanark–Oso meta-anorthosite is the only other unit that lies near this trend. In both figures, the dividing line of Irvine and Baragar (1971) is shown for reference only, as the figure is used for comparative, not classification purposes.

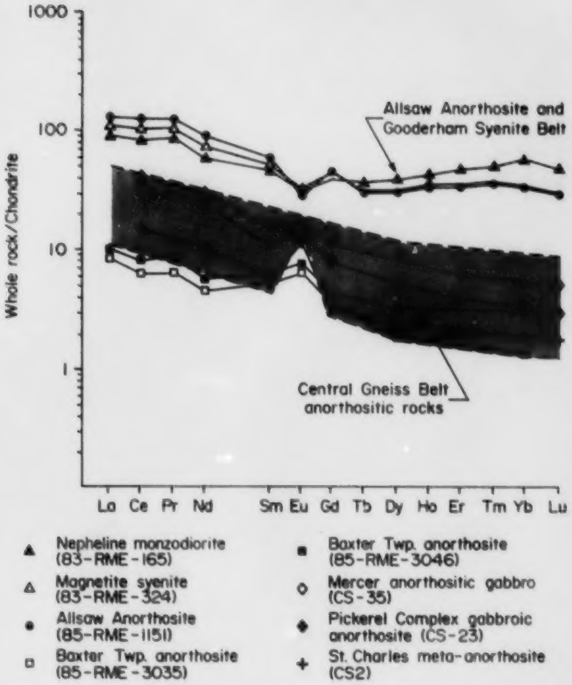


**Figure 19.23.** Chondrite-normalized REE plot of metamorphosed anorthositic rocks from the Grenville Province, Ontario. Note distinct slope and REE abundances of the Allsaw Anorthosite and its similarity to the Gooderham Syenite Belt metamorphosed monzodiorites and syenites. Baxter Township meta-anorthosites (thin anorthosite unit) have REE patterns similar to other Ontario Central Gneiss Belt patterns. Mercer, Pickerel and St. Charles meta-anorthosites have REE patterns within the range of massif-type anorthosites and show a similar spread to Marcy meta-anorthositic rocks (patterned area, data *from* Ashwal and Seifert 1980; Seifert et al. 1977; Simmons and Hanson 1978). Data for samples CS2, CS23 and CS35 are *from* Simmons and Hanson (1978). Other samples are from Easton 1990b.

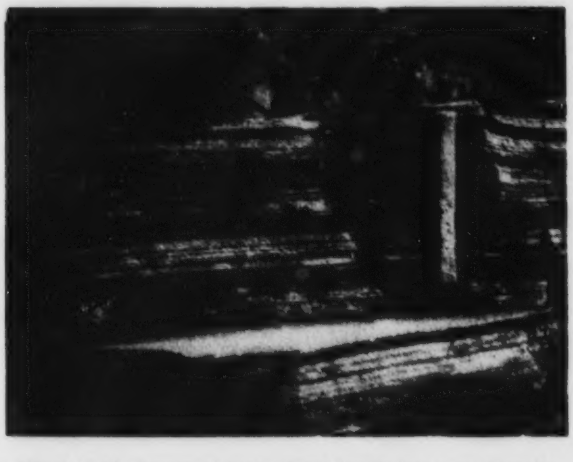


**Figure 19.24.** Transposed gneiss, Parry Sound shear zone, Georgian Bay. Hammer handle is 30 cm long.

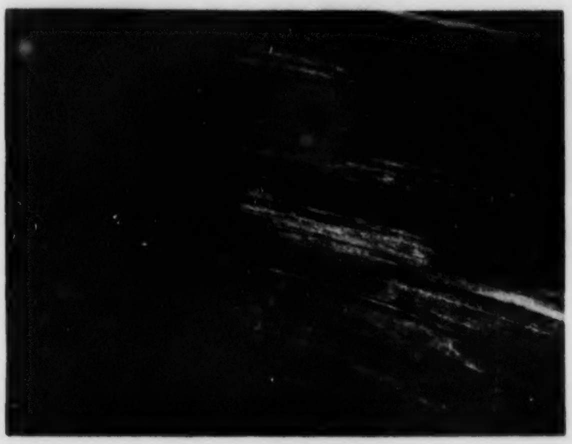


**Figure 19.25.** Straight gneiss, Central Metasedimentary Belt Boundary Zone, near Minden. Hammer handle is 30 cm long.

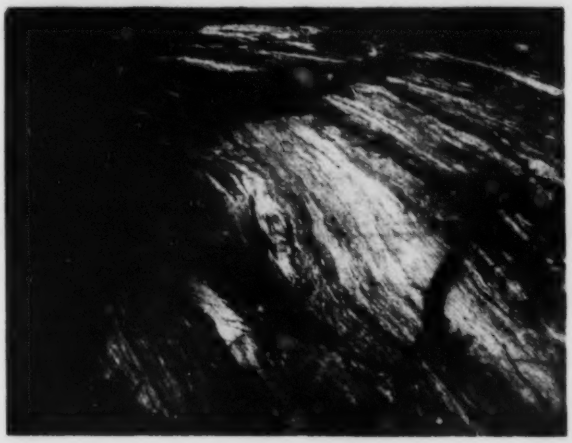


**Figure 19.26.** Irregularly layered and straight gneisses from the Central Metasedimentary Belt Boundary Zone near Minden. Hammer handle (centre right) is 30 cm long.

## PORPHYROCLASTIC GNEISS

A porphyroclastic gneiss consists of a background gneiss of varied composition (quartzofeldspathic to amphibolitic to aluminous) containing isolated potassium feldspar and plagioclase (also locally, garnet and apatite) inclusions, feldspar aggregates and streaks aligned in the foliation. In some outcrops, mechanical and mineralogical disaggregation from pegmatite to feldspar inclusions can be observed (Figures 19.30 and 19.32). Thus, some feldspars are porphyroclasts (Figure 19.31). However, in other instances, some of the feldspars may be porphyroblasts (Figure 19.33). Whatever their origin, isolated feldspar augen are diagnostic of this unit. The groundmass is very fine-grained (*see* Figures 19.30, 19.31, 19.32 and 19.33).

The tectonite units defined above are commonly migmatitic. Migmatization may predate or follow the

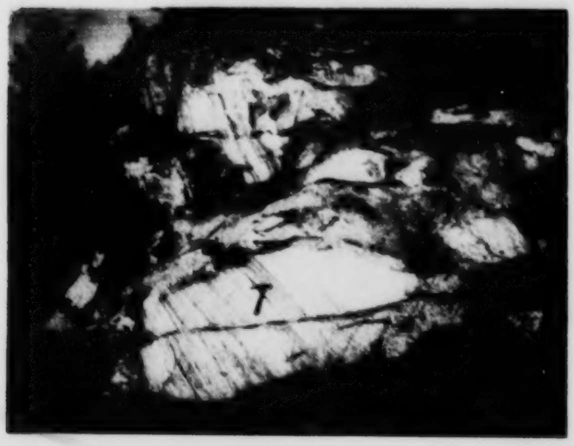


**Figure 19.27.** Block gneiss composed of anorthosite and anorthosite gneiss within a pegmatitic matrix, Highway 518. This zone of anorthositic rocks can be traced for over 30 km. Anorthositic block tectonites are commonly found along terrane and domain boundaries within the Central Gneiss Belt. Hammer handle is 30 cm long.

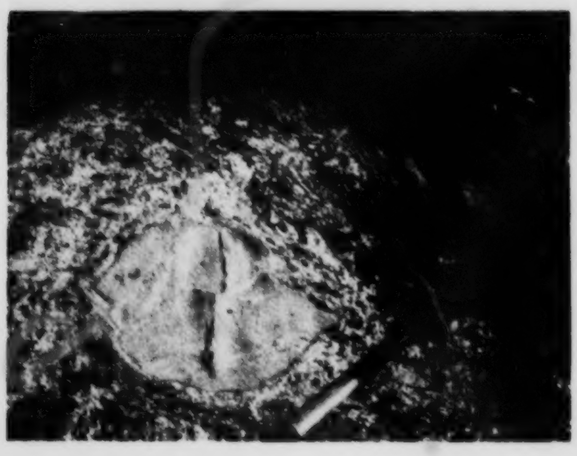


**Figure 19.28.** Block gneiss in granular gneiss, Red Cedar Lake Gneiss, Segment 3 of the Grenville Front Tectonic Zone, Highway 11 north of Marten River. Hammer handle is 30 cm long.

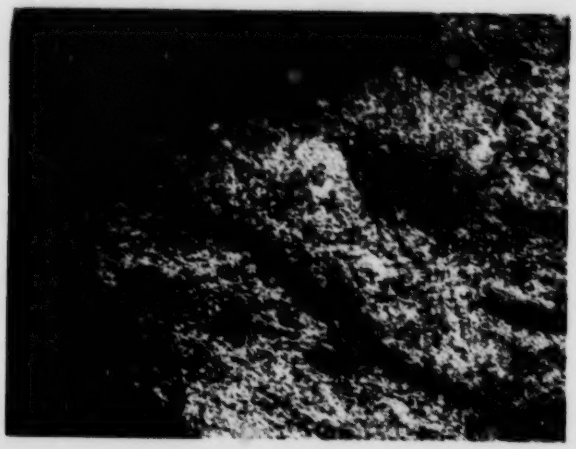


**Figure 19.29.** Granular gneiss, Red Cedar Lake Gneiss, Segment 3 of the Grenville Front Tectonic Zone, southeast of River Valley. Hammer handle is 30 cm long.

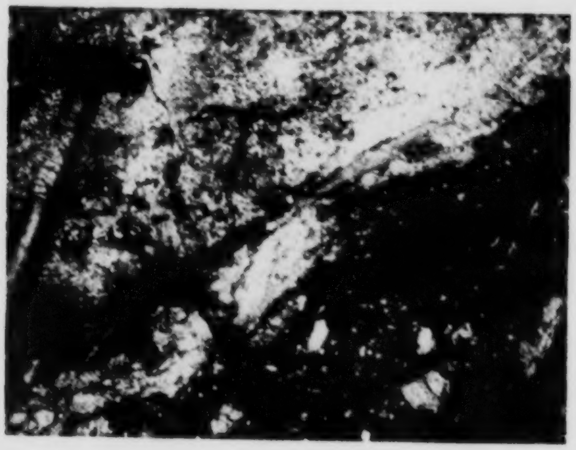


**Figure 19.30.** Porphyroclastic gneiss, Central Metasedimentary Belt, near Minden. In this instance, the feldspar crystals appear to be isolated phenoclasts derived from mylonitization of the pegmatite in the left-hand side of the photo. Hammer handle is 30 cm long.

development of the tectonite. Syntectonic pegmatites are commonly found with these tectonite units and have been useful in dating deformation in these zones.

The tectonites occur at terrane and domain boundaries, where they truncate pre-existing structures and rock units. They also occur internally within domains marking high-strain zones even if there is no lithologic or structural contrast across the high-strain zone (Figure 19.34). The domain-internal tectonite zones are generally narrower and less continuous than at domain-terrane boundaries.

## Protolith Recognition

Repeated metamorphism and deformation makes recognition of protolith problematic in many parts of the Central Gneiss Belt. Metaigneous rocks locally preserve igneous textures in the cores of larger bodies (Figures 19.35, 19.36 and 19.37) and are generally more homogeneous than some of the surrounding gneisses (compare Figures 19.38, 19.39 and 19.40, 19.41, 19.42).This may in part reflect the younger ages of these bodies (ca. 1420 Ma) compared to their gneissic host rocks (ca. 2700 to 1450 Ma).

Recognition of supracrustal protoliths is more difficult. Amphibolitic rocks may have volcanic or plutonic precursors, and in the absence of preserved textures, it may not be possible to distinguish the two. Pelitic metasedimentary rocks are identified from their mineralogy, consisting of mauve garnets, sillimanite or kyanite, abundant biotite, graphite, apatite or diopside. Similarly, marble and quartzite are recognizable even at very high metamorphic grades because of their distinctive composition. It should also be borne in mind that many gneisses in

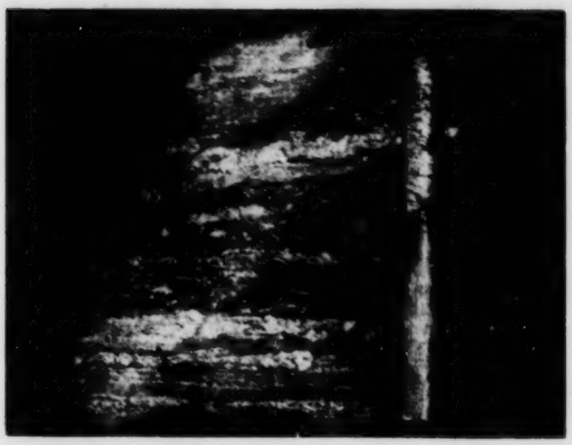


**Figure 19.31.** Porphyroclastic gneiss, Central Metasedimentary Belt, near Minden. Hammer handle is 30 cm long.

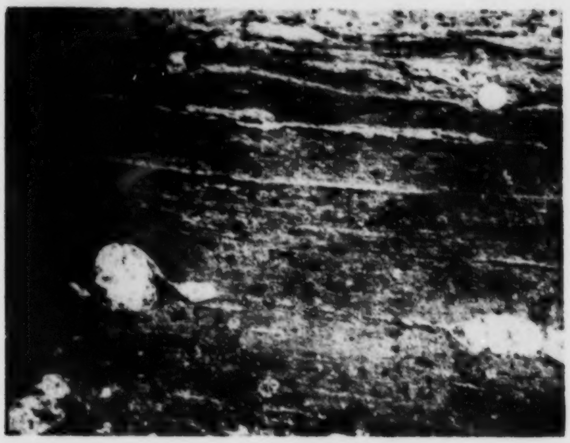


**Figure 19.32.** Rotated feldspar porphyroblasts in fine-grained gneiss. Rotation of feldspar porphyroblasts can be used as sense of shear indicators (*see* Davidson et al. 1982). Coin is 2 cm across.

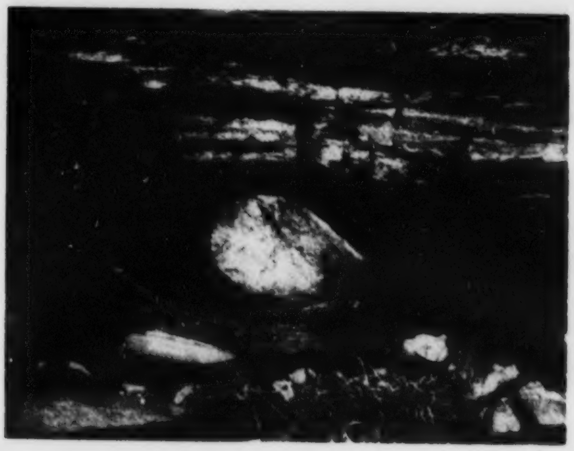


**Figure 19.33.** Large feldspar porphyroblast in straight gneiss in the Central Metasedimentary Belt Boundary Zone, Highway 35, north of Minden. Knife is 9 cm long.

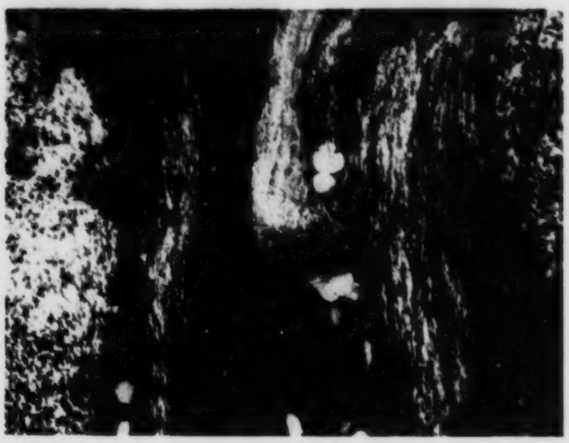


**Figure 19.34.** Shear zone showing progressive deformation from least deformed rocks (left) to most intensely deformed (centre) within the River Valley Anorthosite, south of the Grenville Front, in Segment 2 of the Grenville Front Tectonic Zone. Field of view is 50 cm across.

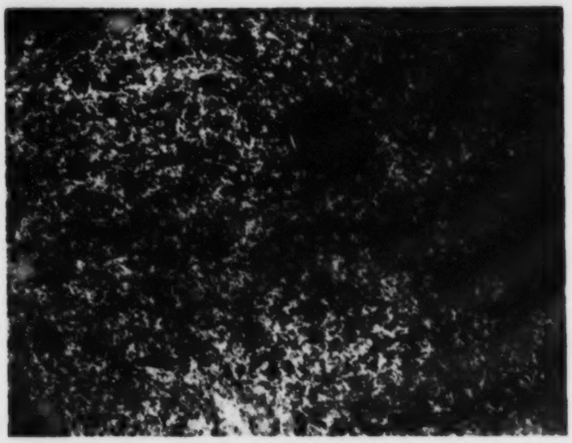


**Figure 19.35.** Recrystallized gabbro from the Parry Sound Domain showing relict gabbroic texture. Lens cap is 6 cm across.

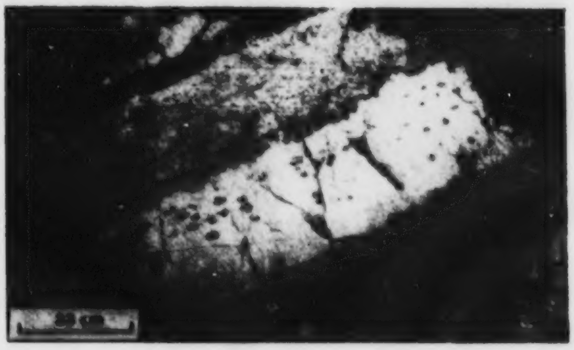


**Figure 19.36.** Igneous layering within anorthosite and gabbroic anorthosite of the Whitestone Gabbroic Anorthosite, Parry Sound Domain. Scale bar is 33 cm long.

**Figure 19.37.** Intrusion breccia within McKellar-type diorite, Parry Sound Domain. Scale card is 9 cm long.

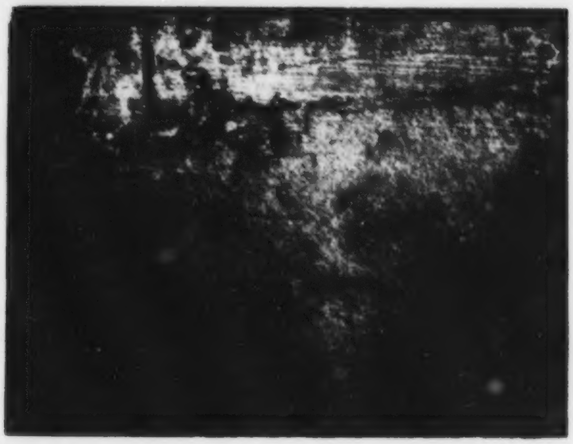


**Figure 19.38.** Migmatitic syenogranite gneiss of the Fishog Domain. Pen is 13 cm long.

**Figure 19.39.** Monzonite gneiss, Fishog Domain. Hammer handle is 30 cm long.

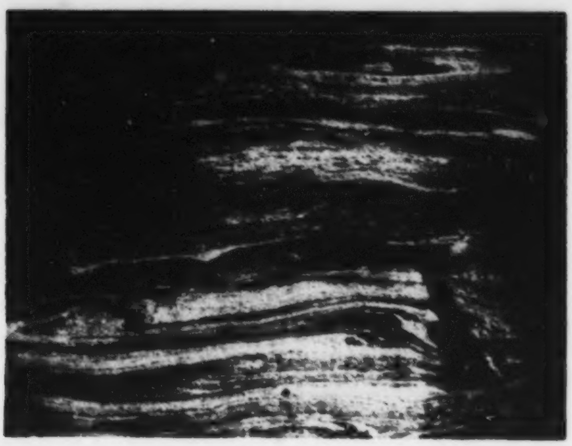


**Figure 19.40.** Typical migmatitic gneisses of the Nepewassi Domain, Nipissing Terrane, near Hagar. Hammer handle is 30 cm long.

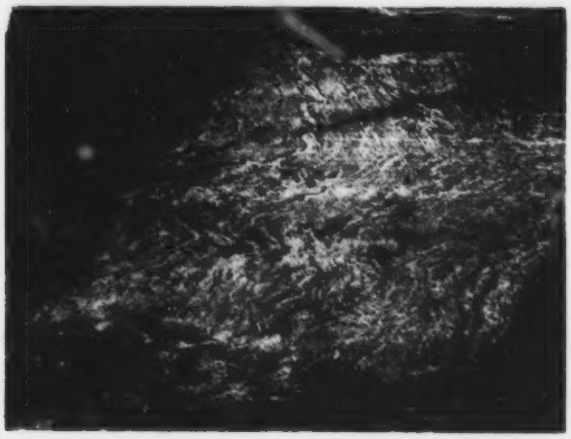


**Figure 19.41.** Typical quartzofeldspathic migmatites of the northern Rosseau Domain, Highway 141. Field of view is 1 m across.

**Figure 19.42.** Typical quartzofeldspathic migmatites of the Britt Domain. Hammer handle is 30 cm long.

the Central Gneiss Belt are polymetamorphic (e.g., in the Nepewassi Domain, metasedimentary gneiss affected by Archean metamorphism have been deformed and metamorphosed once, possibly even twice by the end of the Grenville Orogeny). In such instances, original protolith may be difficult to recognize, and the value of protolith identification in such instances would be questionable.

## METALLOGENY

Mineral exploration within the Central Gneiss Belt and the Grenville Front Tectonic Zone is scant, in part, because of the perception that high-grade gneiss terranes have little mineral potential. Lack of detailed geologic mapping and mineral deposit models for the region have also discouraged exploration. Our understanding of the tectonic makeup of the Central Gneiss Belt allows for the establishment of a broad framework for mineral exploration within the Central Gneiss Belt as detailed by Easton and Fyon (this volume).

Flagstone is probably the only commodity that is not domain specific. This is because the best flagstone prospects are found in transposed and straight gneiss zones associated with terrane-domain boundaries. Recently, exploration of the *circa* 1170 million-year-old suite mafic and ultramafic bodies (*see* Figure 19.20d), also associated with domain-terrane boundaries, for platinum group element, copper and gold mineralization has been undertaken.

Significant metallic and industrial mineral deposits are described under the terrane descriptions below. What follows is an overview of the broader metallogenic characteristics and potential of each terrane summarized from Easton and Fyon (this volume).

The Grenville Front Tectonic Zone is divided into 3 segments reflecting the geology immediately west and north of the Grenville Front (*see* Terrane Descriptions). The lack of closely spaced jointing in many of the metaplutonic rocks in the Grenville Front Tectonic Zone and the Nepewassi Domain accounts for the numerous building stone prospects south of Sudbury (Lacey 1989), especially when compared to the western Central Gneiss Belt region studied by Marmont (1991). The mineral potential of these segments is similar to that present immediately north of the Grenville Front. Segment 1 consists of reworked Killarney Magmatic Belt granitic rocks and has a low metallic mineral potential. Lacey (1989) summarizes the building stone potential of the area.

Segment 2 consists of reworked Southern Province rocks and thus has the same potential as adjacent Southern Province rocks as described by Fyon, Bennett et al. (this volume). Southeast of Wanapitei Lake, the Grenville front apparently truncates gold-bearing, quartz-arsenopyrite- pyrite veins and associated sodium metasomatism. Deformed equivalents of these veins can be expected southeast of the front. Nickel-copper-platinum group element deposits associated with Nipissing diabase intrusions and the East Bull Lake-type gabbro-anorthosite intrusions also may be expected south of the Grenville Front. The River Valley Anorthosite, south of the front, has been quarried for building stone producing an attractive stone called Dana Black. Lacey (1989) also describes several stone prospects from Segment 2.

Segment 3 consists of reworked Superior Province rocks, namely, metasedimentary gneisses south of the Temagami greenstone belt. Migmatitic quartzofeldspathic gneisses in this segment have the same potential as other metasedimentary terranes of the Superior Province, as described by Fyon, Breaks et al. (this volume). The paucity of orthogneiss bodies in this segment of the Grenville Front Tectonic Zone makes it an unfavourable region for stone prospects.

The Nipissing Terrane is underlain by large areas of Archean (Krogh 1989; Dickin and McNutt 1990) migmatitic gneisses. The polymetamorphic character of these rocks prevents any realistic evaluation of their mineral potential, although it is likely to be limited. Paleoproterozoic to Mesoproterozoic supracrustal rocks in the French River area include quartzite and quartzofeldspathic and amphibolitic gneisses and may have the best potential for mineralization in the Nipissing Terrane. Possible mineralization environments to examine in future include placer gold, base metals and iron formation. The building stone potential of the area is reviewed by Lacey (1989), and the area contains several promising prospects.

The Tomiko Terrane consists of a large area of Mesoproterozoic supracrustal rocks. Lumbers (1971b, 1971c) identified several areas of quartz-magnetite iron formation, and the abundance of supracrustal rocks in this terrane suggests a relatively high mineral potential. Several stone quarries are present in the terrane, as are kyanite and vermiculite occurrences (Oswiacki et al. 1989, 1990; Lumbers 1971c).

The best inferences on metallogenic potential of the Algonquin Terrane come from an examination of mineralization types through time and space (e.g., Hutchinson 1981a, 1981b; Veizer et al. 1989). Most Algonquin Terrane rocks were deposited during the Mesoproterozoic to Neoproterozoic (ca. 1.7 to 0.9 Ga) phase of stable cratons (Veizer et al. 1989). Significant ore deposits in this period include unconformity-related uranium, manganese and copper deposits. Other mineral deposits associated with high-grade metamorphic terranes include the Broken Hill lead-zinc deposits, which occur in upper amphibolite to granulite facies, and granulite-facies graphite deposits of the Sri Lankan Shield. Many of the rock types present in the Algonquin Terrane are similar to those associated with the Broken Hill ores, and there is reasonable potential for this type of mineralization within the Algonquin Terrane.

The metasedimentary-rich domains of the Algonquin Terrane host significant graphite deposits; the Cal Graphite Mine in Butt Township began producing early in 1990. Significant graphite deposits have been reported in the Kiosk and Opeongo and Powassan domains (Garland 1987). The regional setting of these deposits is poorly constrained. All occur in granulite facies, quartzose and quartzofeldspathic rocks in high-strain zones near domain boundaries, or in high-strain zones between orthogneiss intrusions. Marmont (1991) summarizes the building stone, flagstone and feldspar resource potential of the Algonquin Terrane.

The abundance of mafic rocks and carbonate and pelitic gneisses in the Parry Sound Terrane makes this region the most extensively prospected, and the best exploration target in the Central Gneiss Belt. Villard et al. (1984) summarize numerous gold prospects in the area. Most are associated with quartz veins, containing associated chalcopyrite, pyrite and pyrrhotite mineralization, and are spatially associated with metamorphosed quartz diorite bodies surrounded by mafic gneisses. Small bodies of massive magnetite occur in Lount Township. The industrial mineral potential of the Parry Sound Terrane is high, and includes alumina (Marmont and Johnston 1987; Marmont 1988a), agricultural lime from marbles (Marmont 1988b), building stone (Marmont 1991) and flagstone (Fouts and Marmont 1989). The most productive flagstone quarries in the Central Gneiss Belt occur in the Parry Sound shear zone.

## Terrane Descriptions

### GRENVILLE FRONT

The Grenville Front (in Ontario, the Grenville Front Boundary Fault) marks the northwest limit of the Grenville Province and has long been recognized as the surface expression of a major crustal discontinuity (*see* Figures 19.3, 19.7a and 19.7b) truncating structures of the geologic provinces to the west and northwest (*see* Figures 19.1 and 19.3). The front is a locus of faulting, mylonitization, metamorphic contrast and locally, truncation of rock units or regional structural trends, or both. It is also expressed by reset potassium-argon and argon-argon isotopic systems (Stockwell 1964; Harper 1967; Easton 1985a, 1986a; York et al. 1991). As an orogenic limit, it has its peculiarities:

1. The adjacent foreland to the northwest does not retain a cover of supracrustal rocks that can be related to the Grenville Orogen, except, possibly, the Seal Lake Group in central Labrador.
2. It does not coincide with the appearance of Grenvillian intrusive age igneous rocks, which occur well to the south.
3. It separates older structural provinces from a broad region of gneisses, some of which are recognizable as their reworked equivalents (e.g., Beaverstone Domain), and others, even though pre-Grenvillian in age, which have no obvious counterparts (e.g., Tomiko Terrane).

Historically, the nature of the Grenville Front in Ontario has been contentious. Initially, it was the eastern limit of Huronian strata (Collins 1925; Quirke and Collins 1930). Gill (1948) noted that the Grenville Front was a locus of faulting, and that a contrast in metamorphic grade occurred across it. J.T. Wilson (1949) labelled this zone the Huron–Mistassini fault zone, and Derry (1950) introduced the term Grenville Front on a tectonic map of Canada without defining its nature. Different opinions were subsequently voiced on whether the Grenville Front represented a fault or a fault zone (McLaughlin 1954), a metamorphic boundary (Hewitt 1956), a "... phenomenon of the draughtsman's table ..." (Engel 1956), or a geochronologic boundary (Wynne-Edwards 1972). Some authors have included rocks northwest of the Grenville Front as part of the Grenville Province (e.g., Stockwell 1964, 1982; Wynne-Edwards 1972).

Where a major fault or mylonite zone marks the Grenville Front, faults, foliations and lineations are commonly steep and inclined to the southeast (Davidson 1986c). It is generally accepted that the Grenville side of the front has been uplifted with respect to the northwest side, based upon metamorphic contrast (Davidson 1986c). Wynne-Edwards (1972) suggested that a foreland belt of deformation, termed the Grenville Foreland Belt, paralleled the Grenville Front as a 30 to 80 km wide zone northwest of the front. This review and Davidson (1986c) found little evidence of a Grenville Foreland Belt.

The nature, significance and duration of tectonic activity along the Grenville Front have been variously interpreted. Gill (1948) suggested that "... movements on low-angle thrust faults carried the Grenville rocks to their present position...." As Davidson (1986c) notes, klippen of Grenville rocks do not occur northwest of the Grenville Front. Lumbers (1971a) argued for compression and thrusting of the Grenville Province against the Southern and Superior provinces, but later suggested (Lumbers 1975) that vertical displacement could account for the features south of the Grenville Front. Wynne-Edwards (1972) showed the Grenville Front as a southeast-dipping thrust fault. With the introduction of plate tectonic theory, the Grenville Front was first considered to be a suture between 2 continental masses (e.g., Dietz 1966; Krogh and Davis 1971b; Gibb et al. 1980). Surface geology does not support the suture hypothesis in that Archean rocks can be traced southeast of the Grenville Front (e.g., Davidson and Bethune 1988; McGrath et al. 1988; Dickin and McNutt 1989a, 1989b, 1990). If any suture is present, it must lie in the core of the orogen (e.g., Rondot 1978; Anderson and Burke 1983), or farther southeast where it has been reworked by the Appalachian Orogen (Dewey and Burke 1973; Baer 1976).

There has also been considerable debate on whether the Grenville Front is a long-lived entity extending back to the Paleoproterozoic (Lumbers 1978; LaTour and Fullagar 1986) or the Neoarchean (Stockwell 1982), or whether it is a feature solely related to the Grenville Orogeny (*see also* summary *in* Davidson 1986c). Deformation and metamorphism of Sudbury Swarm olivine diabase dikes southeast of the Grenville Front (Bethune 1989), implies that the maximum age for Grenville orogenic activity along the Grenville Front is *circa* 1238 Ma (Krogh et al. 1987). If older activity occurred along the Grenville Front, then it is not related to the Grenville Orogeny *senso stricto*.

In Ontario, the Grenville Front can be divided into 3 major segments. Throughout Ontario, the Grenville Front is a fault, termed the Grenville Front Boundary Fault (Card and Lumbers 1977; Lumbers 1978; Davidson 1986c). Changes in the regional trend of the fault occur at the junctions of these segments (Figure 19.43). Studies of the Grenville Front in Ontario are summarized in Table 19.4.

All of the rocks north of the Grenville Front Boundary Fault become increasingly deformed near it. The rocks are overprinted by metamorphism that increases toward the fault, generally reaching lower amphibolite facies just south of the Grenville Front, at least in the River Valley area (Dressler 1979). The front metamorphism is observed in a zone up to 5 km wide in rocks not previously metamorphosed (e.g., Huronian rocks north of segment 3). Dressler (1979), Lumbers (1975) and Davidson (1986c) have noted that Grenville Front-related metamorphism is recognized by the association and orientation of minerals with a southeast-dipping, front-parallel foliation and with downdip lineation. Paleomagnetic poles are reset in a 2 to 5 km wide zone adjacent to the Grenville Front (Stupavsey and Symons 1982a; Hyodo et al. 1986), at least in the Temagami and River Valley areas, congruent with the limited zone of metamorphic resetting associated with the front (Lumbers 1978; Dressler 1979; Davidson 1986c).

In some areas along the Grenville Front, an earlier stage of ductile deformation occurred prior to brittle deformation (Davidson 1986c); however, this ductile deformation need not be related to the Grenville Orogeny. North of the Grenville Front Boundary Fault, rocks are clearly recognizable as being Killarnean, Huronian or Archean, even when metamorphosed and deformed. However, south of the Grenville Front Boundary Fault, protolith is much more difficult to establish, although some distinctive rock units occur south of the boundary fault (e.g., River Valley Anorthosite).

## GRENVILLE FRONT TECTONIC ZONE

The Grenville Front Tectonic Zone (GFTZ) is an ill-defined region up to 30 km across with transitional boundaries that lies mainly south of, but which parallels, the Grenville Front Boundary Fault. The Zone is "... superimposed upon earlier structures in the Grenville, Southern and Superior Provinces..." (Lumbers 1978) and is characterized by northeasterly trending cataclastic foliation and by a prominent southeasterly plunging lineation (Lumbers 1975, 1978). Lumbers (1975, 1978) and Dressler (1979) included rocks north of the Grenville Front Boundary Fault that were overprinted by Grenville metamorphism and containing a southeast-plunging lineation as part of the Grenville Front Tectonic Zone (GFTZ). This usage is retained herein.

Wynne-Edwards (1972) used geophysical criteria, in addition to structure, to characterize the GFTZ, an approach also followed, in part, by Rivers et al. (1989). However, these geophysical characteristics are applicable mainly to the Grenville Front segment in Quebec and, in large part, relate to the greater southward extent of Archean rocks southeast of the Grenville Front in Quebec. In contrast, the

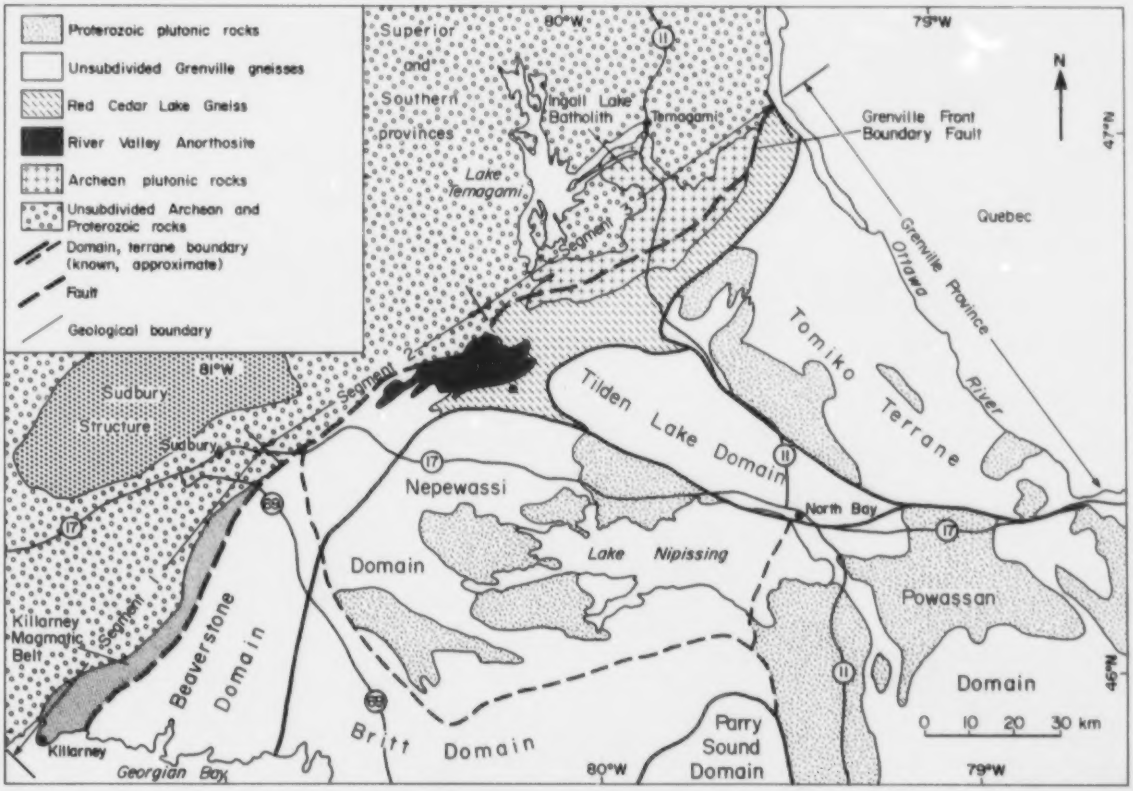

Proterozoic plutonic rocks
Unsubdivided Grenville gneisses
Red Cedar Lake Gneiss
River Valley Anorthosite
Archean plutonic rocks
Unsubdivided Archean and Proterozoic rocks
Domain, terrane boundary (known, approximate)
Fault
Geological boundary
80°W
79°W
81°W
N
47°N
46°N
Superior and Southern provinces
Ingall Lake Batholith
Temagami
Lake Temagami
Segment
Grenville Front Boundary Fault
Quebec
Grenville Province
Ottawa River
Tomiko Terrane
Sudbury Structure
Sudbury
Segment 2
Tilden Lake Domain
North Bay
Nepewassi Domain
Lake Nipissing
Powassan Domain
Killarney Magmatic Belt
Segment
Beaverstone Domain
Britt Domain
Parry Sound Domain
Killarney
Georgian Bay
0 10 20 30 km
11
17
69

Figure 19.43. Geology of the Grenville Front region in Ontario.

**Table 19.4.** Grenville Front studies.

| | |
|---|---|
| | **Segment 1—Killarney to Coniston** |
| **Mapping Studies** | |
| Quirke (1929), Quirke and Collins (1930), Frarey and Cannon (1969), Frarey (1985), Davidson (1986b) | Mapping, includes the Killarney area. |
| Phemister (1960, 1961), Grant et al. (1962) | Mapping studies in the Sudbury area. |
| Culshaw et al. (1988), Davidson and Bethune (1988) | Mapping in the Killarney area focussing on the Georgian Bay shoreline. |
| Davidson (1986b, 1986c) | Review of the Grenville Front in Ontario, Killarney area field trip. |
| Davidson (1992) | Detailed mapping of the Grenville Front near Coniston. |
| **Dike Studies** | |
| Palmer et al. (1977) | Paleomagnetism, K-Ar geochronology and chemistry of the Sudbury Swarm diabase dikes north of the Grenville Front. |
| Bethune (1989), Bethune and Davidson (1988) | Effect of the Grenville Orogeny on the Sudbury Swarm diabase dikes. |
| **Structure and Metamorphism** | |
| Brooks (1964, 1976) | Structural study. |
| Brocum and Dalzeil (1974), Dalziel and Bailey (1968), Dalziel et al. (1969) | Structural and metamorphic studies along the Grenville Front in the Sudbury area. |
| Brown (1968) | Structural study. |
| Clifford (1990) | Structural and geochemical study, Grenville Front near Killarney. |
| Henderson (1967) | Structural study of the Grenville Front near Sudbury. |
| Kwak (1967, 1971) | Metamorphic petrology of rocks along the Grenville Front south of Sudbury. |
| LaTour (1979, 1981a, 1981b), LaTour and Fullagar (1986) | Structural, metamorphic and Rb-Sr isotopic studies of the Grenville Front near Coniston. |
| O'Donnell (1986) | Metamorphic and structural study across the Grenville Front, Killarney area. |
| **Geochronology, Geochemistry** | |
| Dickin et al. (1990) | Nd-isotope study of plutonic rocks along the Grenville Front. |
| Lumbers and Vertolli (1991) | Geochemistry of plutonic rocks along the Grenville Front. |
| Krogh and Davis (1970b), Krogh et al. (1971), Krogh (1989), van Breemen and Davidson (1988a) | U-Pb zircon dating of plutonic rocks in the Killarney Magmatic Belt and south of the Grenville Front. |
| | **Segment 2—Coniston to River Valley** |
| Ashwal and Wooden (1989) | Isotopic study, River Valley Anorthosite. |
| Davidson (1986c) | Review paper on the Grenville Front, field trip in the River Valley area. |
| Dressler (1979) | Detailed mapping of McNish and James townships which straddle the Grenville Front. |
| Lumbers (1971d, 1973) | Mapping. |
| Pearson (1959) | Metamorphic study focussing on kyanite south of the Grenville Front. |
| | **Segment 3—River Valley to Temagami** |
| Barlow (1907) | Reconnaissance mapping. |
| Easton (1988b) | Reconnaissance mapping. |
| Grant (1964) | Rb-Sr study along the Grenville Front. |
| Hyodo et al. (1986) | Paleomagnetic study of the Grenvillian magnetic overprint north of the Grenville Front. |
| Johnston (1954) | Mapping and structural studies. |
| Krogh and Davis (1970b), Steiger and Wasserburg (1969) | Rb-Sr and U-Pb dating along the Grenville Front. |
| Lumbers (1971a, 1971c, 1971d, 1973) | Detailed mapping. |
| Lumbers (1978) | Summary of mapping studies, field trip guide to the Grenville Front Tectonic Zone in the River Valley and Temagami areas. |
| York et al. (1991) | Ar-Ar study across the Grenville Front along Highway 11. |

GFTZ in Ontario is poorly defined geophysically in Ontario (*see* Geophysics).

In Ontario, however, in addition to a purely structural definition, the GFTZ can be defined as a zone where metamorphosed rocks of the Superior or Southern Province can be traced across the Grenville Front into the Grenville Province. Lithologic changes can be mapped out along strike within the GFTZ, coincident with changes in the character of the rock units present northwest of the front, and consequently these gneisses are probably, for the most part, autochthonous. Farther south are found distinctive lithotectonic domains that are parautochthonous (e.g., Nipissing Terrane) or allochthonous (e.g., Tomiko Terrane). Recent work (Haggart et al. 1992; Culshaw et al. 1992) indicates that the Grenville Front Tectonic Zone may have migrated northward during the Grenville Orogeny. In the Pickerel River area, Culshaw et al. (1992) report overprinting of GFTZ style deformation fabrics by "Grenvillian" fabrics. Thus, the GFTZ may represent the final position of a northward migrating crustal ramp related to the Grenville Orogeny.

The GFTZ can be divided into 3 broad lithologic segments corresponding to the 3 segments of the Grenville Front Boundary Fault (*see* Figure 19.43).

## Segment 1: Killarney–Sudbury Segment Including the Beaverstone Domain

This area has been mapped by Lumbers (1975), Frarey (1985), Davidson and Bethune (1988) and Culshaw et al. (1988), with additional studies being listed in Table 19.4. The divergence of opinion regarding the placement of the Grenville Front in the Killarney area was discussed under the section on the Killarney Magmatic Belt. The placement of the Grenville Front Boundary Fault in the area follows that of Ayres et al. (1971a), Lumbers (1975) and Davidson (1986b, 1986c).

Near Killarney, plutonic rocks of the Killarney Magmatic Belt can be traced east and southeast of the Grenville Front Boundary Fault (Davidson 1986c; Davidson and Bethune 1988; Culshaw et al. 1988; Frarey 1985; *see* Figures 19.3 and 19.43). As discussed under the section on geophysics, the distinctive aeromagnetic and gravity signature of the Killarney Magmatic Belt can be traced along the north shore of Lake Huron until just west of the Pickerel Complex (*see* Figures 19.3 and 19.11). Further, metamorphosed equivalents of Sudbury Swarm diabase dikes occur throughout this region (Bethune and Davidson 1988; Bethune 1989). This distinctive terrane is termed the Beaverstone Domain (Beaverstone Terrane *of* Rivers et al. 1989). The eastern limit of the Beaverstone Domain is defined by a major mylonite zone located at the main outlet of the French River (Davidson and Bethune 1988). Although located in the Grenville Front Tectonic Zone, the Beaverstone Domain is structurally and lithologically distinct. Other segments of the Grenville Front Tectonic Zone are insufficiently characterized to subdivide into domains.

A variety of paragneisses occur within the Beaverstone Domain (Frarey 1985). Most abundant are migmatitic biotite paragneisses, commonly layered, locally containing graphite and up to 20% garnet and sillimanite. Interlayered with these biotite paragneisses are metaquartzite units, which were correlated, across the Grenville Front, with the Lorrain or Bar River formations of the Huronian Supergroup (e.g., Quirke and Collins 1930; Brooks 1967; Frarey 1985). Krogh (1989) analyzed zircons from one of these metaquartzite horizons on Beaverstone Bay and found 2 populations, one of Archean age and the other roughly 1700 million years in age. If the later zircon population is indeed detrital, then the quartzites are likely derived from the Killarney Magmatic Belt, and not correlative with the Huronian Supergroup. Metaquartzite and metasedimentary rocks from the Tomiko Terrane and the French River area also have *circa* 1700 million-year-old detrital zircon populations, suggesting, perhaps, a widespread period of platformal sedimentation following accretion of the Killarney Magmatic Belt to North America.

The Beaverstone Domain is a wedge tapering northward toward Coniston, where the easternmost Killarney-type granitoid has been identified (LaTour 1979; LaTour and Fullagar 1986; *see* Figure 19.3). The Killarney Magmatic Belt contains plutons of both 1740 to 1720 and 1500 to 1450 million year age. Zircons and titanites from leucosome pods in the Killarney area 12 km south of the front yield an age of 1453±7 Ma for the migmatitic metamorphism (Krogh 1989). This age is comparable with the main period of granite magmatism in the Eastern Granite–Rhyolite Province and is coincident with a period of granulite-facies metamorphism in the Britt Domain (Culshaw et al. 1991; Ketchum et al. 1992). U-Pb titanite determinations and zircon system lower intercepts may be interpreted to indicate that Grenville metamorphic overprinting in the area occurred at *circa* 995 Ma (Krogh 1989; Haggart et al. 1992). It is this event that metamorphosed Sudbury Swarm dikes in the Beaverstone Domain.

No major mineral deposits or mines are known from the Beaverstone Domain. Lacey (1989) examined several plutons in the Killarney Magmatic Belt and the Beaverstone Domain for their building stone potential.

## Segment 2: Sudbury–River Valley Segment

The eastern part of the GFTZ is divided into 2 segments primarily on the basis of variations in the abundance of mafic metaigneous rocks and structural trends. In part, these differences may reflect the geology immediately north of the front.

Two sets of mafic metaigneous rocks are present in this segment. Metamorphosed gabbroic anorthosite and anorthosite of the *circa* 2490 million-year-old River Valley Anorthosite (Ashwal and Wooden 1989) and gabbroic anorthosite gneiss, probably tectonic slivers from the River Valley body, occur throughout the Grenville Front Tectonic Zone to the southwest. In addition, mafic metaigneous rocks in the segment increase in abundance to the east and may represent deformed equivalents of the *circa* 2219 million-year-old Nipissing Diabase Suite (e.g., Dressler 1979).

Equivalents of the Red Cedar Lake Gneiss (*see* Segment 3) are present in segment 2, but they become discontinuous to the southwest due to increased reworking of the Paleoproterozoic crust.

The River Valley Anorthosite has been quarried for building stone in a little recrystallized part of the body, producing a distinctive black stone (Lacey 1989). Although a variety of copper, nickel and gold occurrences occur within the Southern Province (e.g., Card and Lumbers 1977; *see* Fyon, Bennett et al., this volume), no occurrences have yet been reported in this segment of the Grenville Front Tectonic Zone.

### Segment 3: River Valley–Temagami Segment

The Grenville Front Boundary Fault in this segment cuts across the *circa* 2737 million-year-old Ingall Lake Granodiorite (Ar-Ar hornblende age, York et al. 1991) of the Temagami greenstone belt. South of the front, the granodiorite is finer grained, more attenuated and more gneissic than to the north, but it can be traced at least 2 km south into the Grenville Front Tectonic Zone (Lumbers 1971b, 1971c). Another metaigneous body, just north of Marten River, was dated by Rb-Sr whole-rock methods at 2643 Ma by Krogh and Hurley (1968). The paucity of intrusions in this segment is in marked contrast to segment 2.

In the River Valley and Tomiko areas, Lumbers (1971b, 1971c, 1973, 1978) defined the Red Cedar Lake Formation. This unit occurs as a band, up to 15 km wide (*see* Figure 19.43), that lies immediately south of, and parallel to, the Grenville Front from River Valley to the Ottawa River. Lumbers (1978) interpreted this unit as forming from a Paleoproterozoic, turbidite fan system that developed on Archean basement at the edge of the Superior Province, that was subsequently metamorphosed and deformed during the Grenville Orogeny. The unit has been traced in Quebec, where Davidson (1986c) has suggested it represents Pontiac Subprovince (*see* Jackson and Fyon, this volume) gneisses that were reworked during the Grenville Orogeny.

Easton (1988b) redefined the unit as the Red Cedar Lake Gneiss, emphasizing its metamorphic character and established reference sections on Red Cedar Lake, on Highway 805 and the CPR rail-line 0.5 km south of River Valley and on Highway 11 between Marten River and Opechee Creek. The Red Cedar Lake Gneiss is predominantly a granular gneiss (*see* Figures 19.28 and 19.29). Locally it is a deformed augen gneiss (*see* Figure 19.29), a block tectonite or a porphyroclastic gneiss. Locally it may be derived from deformed metaigneous rocks of Archean? age. Much of the tectonism preserved in this gneiss appears to be older than the main Grenville metamorphism and deformation. Krogh (1989) dated monazite from a postmigmatization pegmatite within the Red Cedar Lake Gneiss at 2642±2.4 Ma, with a lower intercept of 992±4 Ma. Steiger and Wasserburg (1969) also reported a U-Pb age of 2647±12 Ma from a leucosome pod from the Red Cedar Lake Gneiss near River Valley. These data suggest Archean deformation and migmatization.

No major mineral deposits have been reported from this segment of the Grenville Front Tectonic Zone.

### Discussion

If the Grenville Front Tectonic Zone does, in fact, contain rock types present immediately to the north, some of which represent either cover sequences or relatively high level intrusions, then only a relatively minor component of vertical movement has occurred along the front. This does not, however, eliminate the possibility that there was considerable structural relief across the boundary much of which has been subsequently removed.

## NIPISSING TERRANE

### Nepewassi Domain

#### GEOLOGY

The Nepewassi Domain (Easton 1988b) was discriminated from its neighbours on the basis of structural trends as well as rock types. The area underlain by the Nepewassi Domain has been mapped by Lumbers (1971a, 1973 and 1975) at 1:63 360 scale. New aeromagnetic maps (*see* Maps 2586 and 2587, map case) show a well-defined northwestern margin to the domain even though the Nipissing Terrane has a subdued magnetic pattern.

The Nepewassi Domain is underlain by compositionally heterogeneous migmatitic (*sensu stricto*) gneisses (*see* Figure 19.40) which probably have a polycyclic history. Near Hagar, these gneisses (*see* Figure 19.40) are 2737 Ma (Krogh 1989) but furnish titanite ages of 2727 Ma indicating Archean metamorphism. This is, however, about 80 million years older than Archean metamorphism dated from the Grenville Front Tectonic Zone to the northeast (*see* discussion of Red Cedar Lake Gneiss *in* Segment 3), suggesting that the Archean rocks in the Nepewassi Domain may be different from those a few kilometres to the north in the Grenville Front Tectonic Zone. Paragneisses and orthogneisses from the Nepewassi Domain all show Archean or Paleoproterozoic Nd-Sm model ages (Dickin and McNutt 1990; Dickin et al. 1990). Plutonic rocks in the domain form 2 suites which are less deformed than their migmatitic host rock: an older, *circa* 1700 million-year-old (Lumbers 1975) granite-monzogranite suite which includes the West Bay and Sturgeon Falls granites and a younger, *circa* 1450 to 1420 million-year-old suite which includes the Cosby Monzonite. Locally, igneous contacts are preserved, and relict igneous textures are commonly present in these plutons. Anorthosite bodies of the St. Charles and Mercer meta-anorthosites cut the West Bay Granite and may be as young as 1222 million years (Prevec 1992). Chemical analyses of the plutonic rocks are presented in Lumbers (1975).

Quartz-feldspar paragneisses, interlayered with minor amphibolite in the French River area, form a large, southeast-plunging synclinorium intruded by the Cosby Monzonite. The paragneisses contain zircons indicating a single provenance source of 1744±11 Ma. Grenville-age leucosome pods in the same gneisses have been dated at 995 Ma (Krogh 1989; Krogh and Davis 1969, 1970a, 1971a). Zircons from a nearby quartzite horizon, however, contained only zircons of Paleoproterozoic and Archean ages (Krogh 1989), indicating a different source. The presence of Archean migmatites and Mesoproterozoic paragneisses in the Nepewassi Domain suggest that in the French River area we are seeing a Mesoproterozoic metasedimentary sequence developed upon an Archean or Paleoproterozoic basement, both of which were subsequently deformed during the Grenville Orogeny obscuring this early basement-cover relationship.

#### MINERAL DEPOSITS

No significant mineral deposits have been reported from the Nepewassi Domain. Alkalic complexes of Early Paleozoic age in the Lake Nipissing area have potential for iron and niobium (Lumbers 1971; Currie 1976). Lacey (1989) describes several potential building stone prospects in the domain. Jointing is well-spaced and moderate within the Nepewassi Domain compared to the Algonquin Terrane, making the Nepewassi Domain particularly attractive for prospecting for building stone.

### Tilden Lake Domain

#### GEOLOGY

The Tilden Lake Domain was defined on structural and lithologic characteristics by Easton (1988b). Most of the domain has been mapped at 1:63 360 scale by Lumbers (1971a, 1971b, 1971c, 1973); it contains heterogeneous migmatitic gneisses of unknown protolith similar to those found in the Nepewassi Domain. Unlike the Nepewassi Domain, metaigneous rocks of both the *circa* 1700 and *circa* 1450 million-year-old suites are uncommon in this terrane. Paragneiss is more abundant in the Tilden Lake Domain than in the Nepewassi Domain, however, Mesoproterozoic paragneisses similar to the French River gneisses are absent. Nd-Sm model ages from the Tilden Lake Domain show Archean or Paleoproterozoic ages (Dickin and McNutt 1989a, Dickin et al. 1990).

#### MINERAL DEPOSITS

No major mineral deposits have been reported from this domain.

## TOMIKO TERRANE

#### GEOLOGY

The Tomiko Terrane (*see* Figure 19.2) was originally defined by Easton (1988b) on the basis of: 1) regional structural trends; 2) an abundance of metasedimentary rocks compared to other domains within the Nipissing area; and 3) mapping along the western boundary of the terrane of shear zones similar to those present along other domain boundaries within the Central Gneiss Belt.

A limited number of Mesoproterozoic Nd-Sm model ages (Dickin and McNutt 1989a, 1989b; Dickin et al. 1990) of paragneisses from the Tomiko Terrane contrast with Archean and Paleoproterozoic model ages of the adjacent Nipissing Terrane and the Grenville Front Tectonic Zone. Data from metaigneous rocks from the Tomiko Terrane, except the Mulock Granite, (Dickin et al. 1990) also give Mesoproterozoic ages.

Metaigneous rocks in the Tomiko Terrane are granitic in composition (in contrast to monzonitic plutons in the Algonquin Terrane) and differ in composition from the *circa* 1.7 billion-year-old granitic plutons in the Nipissing Terrane (Lumbers 1971a–d). The largest of these plutons is the Mulock Granite, dated at $1244^{+4}_{-3}$ Ma (Lumbers et al. 1991), the only known pluton of this age in the Central Gneiss Belt. The Mulock Granite has an exceedingly old Nd-model age of 2440 Ma (Dickin et al. 1990), suggesting that the Mulock Granite was derived from melting of Paleoproterozoic crust without any juvenile mantle component. Krogh and Davis (1969) reported a *circa* 1200 million-year-old Rb-Sr age for the Mulock Granite, suggesting that Rb-Sr isotopic resetting in the Tomiko Terrane may be limited. Baer (1980) reported a Rb-Sr date of 1137±173 Ma from the Timber Lake (Jocko) granite, similar to the age of the Mulock Granite.

Dating of detrital zircons from the metamorphosed, extensive Mattawa Quartzite in the Tomiko Terrane by Krogh (1989) indicates that the metaquartzite exclusively contains Mesoproterozoic zircons (1687±20 Ma), despite the current proximity of the terrane to the Superior Province. In contrast, metaquartzites from the French River and Killarney areas, and paragneisses from near Hagar, have mixed zircon populations of Archean and Paleo- to Mesoproterozoic age, or Archean ages, exclusively (Krogh 1989). The Tomiko Terrane is a Mesoproterozoic terrane that is probably allochthonous with respect to the reworked Archean and Paleoproterozoic rocks of the GFTZ and the Nipissing Terrane. Evidence to suggest this is: 1) the presence of local concentrations of iron formation (Lumbers 1971b, 1971c) unique in this segment of the Central Gneiss Belt; 2) the distinct zircon provenance of the metaquartzites; and 3) the neodymium-samarium data.

Peak metamorphic temperatures and pressures in the Tomiko Terrane are notably lower (less than 700°C and 6 to 8 kb) than those from the Nipissing Terrane (greater than 700°C and 8 to 10 kb) and the Algonquin Terrane to the south (Annovitz and Essene 1990). They attributed the difference in metamorphic pressures between the Tomiko and Algonquin terranes to vertical movement along Ottawa–Bonnechere Graben faults parallel to the Mattawa River. However, this required movement the reverse of that observed by Lumbers (1971a). An alternative explanation is that the difference in metamorphic conditions reflects the allochthonous nature of the Tomiko Terrane, with the difference in grade related to stacking of a less deeply buried terrane atop the parautochthon, or a metamorphic history different from the adjacent terranes, or both. The west to east variation in P-T conditions between the Nipissing and Tomiko terranes noted by Anovitz and Essene (1990) is in general agreement with the proposed terrane boundary.

#### MINERAL DEPOSITS

No major mineral occurrences have been reported from the Tomiko Terrane. Lumbers (1971c) mapped quartz-magnetite iron formation in the northern part of the terrane, and metamorphosed iron formation was reported from Hammell Township (Owsiacki et al. 1990). Exploration has also been conducted in Notman and Hammell townships for base metals and gold and platinum in gabbroic intrusions, on the assumption that metamorphosed equivalents of the Archean Temagami greenstone belt underlie this region (Owsiacki et al. 1989). Titaniferous-iron deposits are associated with the Fanny Lake gabbro body in the northern part of the terrane (Lumbers 1971b; Owsiacki et al. 1989).

Kyanite and pegmatite occurrences in the Tomiko Terrane have also been examined for industrial mineral potential (Owsiacki et al. 1989). Lumbers (1971c) reported several vermiculite occurrences associated with calc-silicate gneisses in Butler Township that may be related to pre-Paleozoic weathering associated with the proximity of the Precambrian–Paleozoic unconformity surface along the Timiskaming–Ottawa graben. Feldspar and mica were produced from the Powassan Domain, adjacent to the southern boundary of the Tomiko Terrane in the 1930s and 1940s in Mattawan and Olrig townships (Harding 1946). Several stone quarries operate in the north half of the terrane (Oswiacki et al. 1989, 1990).

## ALGONQUIN TERRANE

### Britt Domain

#### GEOLOGY

Davidson and Morgan (1981), Davidson et al. (1982) and Culshaw et al. (1988, 1989, 1991) describe the Britt Domain. Figure 19.44 shows the major units within the domain. Of all the domains within the Algonquin Terrane, the Britt Domain is the most intensely studied.

The Britt Domain consists of highly deformed gneisses and migmatites of supracrustal and igneous origins (ca. 1680 Ma) that have been intruded by a suite of younger *circa* 1450 million-year-old (van Breemen et al. 1986) plutons, followed by further migmatization and deformation. The prominent structural pattern is outlined by the younger plutonic rocks, now sheet-like masses that delineate northwest-trending, large-scale folds whose axes plunge mainly to the southeast (*see* Figure 19.44).

#### GNEISS ASSOCIATIONS

Culshaw et al. (1988, 1989, 1990, 1991) have divided the Britt Domain into a number of "gneiss associations." Although the term is primarily lithologic, individual or groups of gneiss associations may display differences in magmatism, metamorphism or mafic diking. These differences form the basis for the field-based tectonic interpretation (*see* Figure 19.45; Culshaw et al. 1991). In addition, the terms "monocyclic" and "polycyclic" are used here to distinguish rocks units (not regions, as *in* Rivers et al. 1989) that have undergone, respectively, one or more than one orogenic event. These terms are useful where polycyclic gneiss associations are intruded by monocyclic (single cycle) plutons.

There are 5 gneiss associations in the Britt Domain (*see* Figure 19.44). The descriptions below are based on Culshaw et al. (1991).

#### Key Harbour Gneiss Association

At Key Harbour, the polycyclic Key Harbour gneiss association (*see* Figure 19.44) forms roughly half of the area, the rest being monocyclic igneous rocks. It is largely composed of very leucocratic, pink to grey, quartzofeldspathic orthogneisses and paragneisses (at sillimanite grade), which were variably deformed prior to the emplacement of the *circa* 1684 million-year-old (Corrigan 1990) polycyclic Key Harbour leucogranite. The gneiss is accompanied by subunits of felsic gneiss, calc-silicate gneiss, quartzite and garnet amphibolite. Leucocratic orthogneiss (Key Harbour leucogranite) is weakly migmatitic and discordant to the paragneiss. Other units (Culshaw et al. 1988) include the monocyclic? gabbro-tonalite to granodiorite-granite Free Drinks gneiss and the layered Still River gneiss, a strongly tectonically modified assemblage.

Monocyclic plutons of *circa* 1450 million year age (Britt Granodiorite, Mann Island Granodiorite and the Pickerel Complex *of* Davidson et al. 1982) form extensive sheets defining regional scale folds (*see* Figure 19.44). Despite the fact that they have been moderately to extensively deformed and metamorphosed, they can be readily recognized due to their compositional homogeneity and local preservation of relict igneous textures.

The Mann Island Granodiorite forms grey, elongated sheets which truncate the country rock foliation at a low angle. It is a variably migmatitic, potassium feldspar megacrystic, hornblende-biotite granodiorite as is the Britt Granodiorite (Davidson et al. 1982).

The Pickerel Complex (Quirke 1929; Hewitt 1961, 1962; Lumbers 1975; Davidson et al. 1982; Davidson and Bethune 1988) is characterized by migmatitic hornblende monzogranite with minor quartz syenite and has a central zone of meta-anorthositic and metagabbroic gneiss. The complex is folded within a large synform (*see* Figure 19.44), the Key Harbour synform. Hewitt (1962) reports a quartzite horizon marking the outer limb of the synform. Chemical analyses of Pickerel Complex rocks are found in Lumbers (1975).

In the north, the Rutter Monzonite is in fault contact with the Bigwood Alkalic Complex, which includes 2 lenses of nepheline syenite separated by alkalic syenite (Hewitt 1961). Hewitt (1961) provides chemical analyses of the Bigwood Alkalic Complex rocks. Hewitt (1961) concluded that the Bigwood Alkalic Complex was emplaced as a sill-like body into the country rocks along a north-striking fault zone that parallels the regional structure. Lumbers (1975) reports field evidence indicating the Bigwood Alkalic Complex is younger than the *circa* 1450 million-year-old plutons in the area, consistent with Hewitt's (1961) observations. Lumbers (1975) noted that the Bigwood Alkalic Complex was similar to alkalic complexes in the Bancroft Terrane nepheline belts. If so, the Bigwood Alkalic Complex may represent *circa* 1240 to 1220 million-year-old magmatism within the Central Gneiss Belt. Lumbers (1976c) reports nepheline syenite in spatial association with the Eau Claire meta-anorthosite in the Powassan Domain near Mattawa, which may be of similar relative age to the Bigwood Alkalic Complex.

Numerous crosscutting metamorphosed mafic dikes postdate the monocyclic granitoids, and in polycyclic gneisses they postdate relicts of granulite-facies metamorphism. Some may be related to the Sudbury dike swarm (Davidson and Bethune 1988; Bethune 1989).

### Bayfield Gneiss Association

The Bayfield gneiss association (*see* Figure 19.44) is composed predominantly of *circa* 1680 million year old granitoid orthogneiss. The principal members are grey hornblende-bearing granodioritic to tonalitic orthogneiss and pink biotite granite. Mappable units of pink leucogneiss occur within the orthogneiss and also appear as small crosscutting bodies. Pelitic gneiss is minor. There are several probable monocyclic gabbro plutons and 2 sets of metamorphosed mafic dikes crosscutting pre-Grenvillian leucosomes throughout the Bayfield gneiss association.

### Nadeau Island Gneiss Association

The Nadeau Island gneiss association (*see* Figure 19.44) is characterized by a leucocratic granitic to granodioritic orthogneiss intermixed with grey granodiorite to tonalite orthogneiss and a high proportion of metasedimentary rocks. The *circa* 1680 million-year-old orthogneiss members resemble the Bayfield gneiss association. The metasedimentary rocks comprise a variety of pelitic and semipelitic types, including corundum-bearing varieties. The association contains evidence of a pre-Grenvillian upper amphibolite- and granulite-facies metamorphism. The Nadeau Island gneiss association contains the same crosscutting mafic dike suites as the Bayfield gneiss association. A paragneiss sample from the Pointe au Baril area, analyzed by Rb-Sr whole-rock methods, yielded an age of 1752±95 Ma (Krogh and Davis 1969, 1970a, 1970b, 1971b).

The Nadeau Island gneiss association is bounded along its upper and lower margins by extensive *circa* 1450 million-year-old monocyclic granitoids, the Shawanaga Granodiorite and the Pointe au Baril suite (*see* Figure 19.44). The Pointe au Baril suite varies from granite, through granodiorite to mafic tonalite. Originally mapped by Davidson et al. (1982), these monocyclic plutons resemble their monocyclic counterparts in the Key Harbour gneiss association (Culshaw et al. 1991).

### Ojibway Gneiss Association

The metaplutonic Ojibway gneiss association (*see* Figure 19.44) lacks crosscutting mafic dikes, is not cut by monocyclic granitoids and contains no evidence for pre-Grenvillian metamorphism. Thus, Culshaw et al. (1991) suggest an age of *circa* 1450 Ma or less for this gneiss association. The dominant unit is a grey granodioritic-tonalitic orthogneiss. Grenvillian migmatization becomes progressively more intense at higher structural levels to the southeast (*see* Figure 19.44). The Ojibway gneiss association contains globular and pod-like mafic bodies including coronitic metagabbro, garnet- clinopyroxene-bearing metabasites (pseudoeclogites) and minor anorthosite (cf., Needham 1987; Davidson et al. 1982). Many of these are concentrated close to the boundary with the Nadeau Island gneiss association.

The boundary with the underlying Nadeau Island gneiss association is sharp and strongly deformed. On account of this, and because it separates units with contrasting histories, Culshaw et al. (1991) interpret it as a cryptic thrust.

### Sand Bay Gneiss Association

The Sand Bay gneiss association (*see* Figure 19.44) is largely of supracrustal origin, probably younger than *circa* 1450 Ma (Culshaw et al. 1991, 1992) and includes quartzofeldspathic migmatites, quartz-plagioclase-biotite-muscovite (± amphibole, ± epidote) paragneiss (Dillon Schist), metaquartzite, pink leucocratic gneiss, calcareous gneiss, rare marble and substantial units of hornblende gneiss and amphibolite. Like the Ojibway gneiss, with which it is thought to form a single infolded tectonic unit, it lacks mafic dikes, monocyclic granitoids and relicts of older high-grade metamorphic assemblages.

The Sand Bay gneiss association is overlain immediately to the southeast by a unit of grey gneiss of mixed protolith located at the base of the Parry Sound shear zone. The boundary of this unit with the Sand Bay gneiss association is marked in the east by the *circa* 1346 million-year-old monocyclic orthogneiss (Nobel Granodiorite or "marginal orthogneiss" of van Breemen et al. 1986).

## STRUCTURE

The principal structures in the Britt Domain are regional-scale, northwest-trending, horizontal upright folds extending from the southern margin of the Grenville Front Tectonic Zone to the Parry Sound Domain (Davidson et al. 1982; Schwerdtner 1987; *see* Figure 19.44). The folds formed during progressive Grenvillian metamorphism with axes paralleling a regionally penetrative stretching lineation. The stretching lineation, with important exceptions, generally plunges gently to the southeast. Minor folds, including sheath folds, have axes parallel to this lineation. This lineation marks the orogenic transport direction, the same fabric occurs in the Grenville Front Tectonic Zone (Davidson and Bethune 1988) and in the Central Metasedimentary Belt Boundary Zone (Culshaw et al. 1983; Hanmer 1988).

The folding is accompanied by the formation of a planar-linear (L-S) fabric in the monocyclic plutonic rocks and the transposition of earlier foliations in the polycyclic gneisses. Within low strain zones, pre-Grenvillian fabrics can be recognized where they are cut by pre-Grenvillian mafic dikes, which, themselves, preserve early-Grenvillian fabrics (Culshaw et al. 1991).

South of Key Harbour are 2 belts of mylonitic gneiss which developed during progressive Grenvillian metamorphism (*see* Figure 19.44) and are coeval with the formation of the northwest-trending folds. The largest belt is the regionally extensive Central Britt shear zone (CBSZ) (*see* Figure 19.44). Most kinematic indicators within the

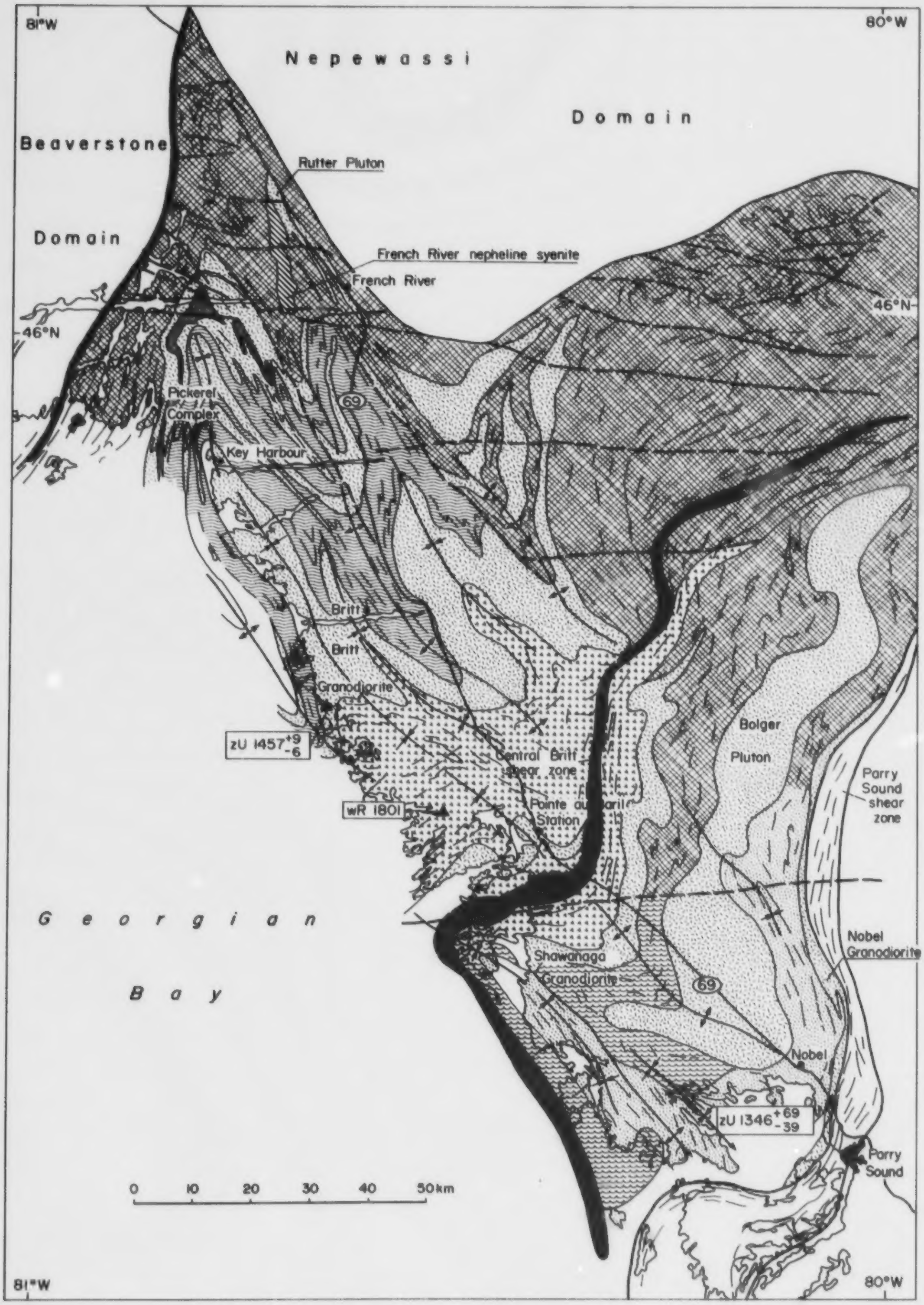

81°W
80°W
Nepewassi
Domain
Beaverstone
Domain
Rutter Pluton
French River nepheline syenite
French River
46°N
46°N
Pickerel Complex
Key Harbour
69
Britt
Britt Granodiorite
zU 1457 +9 -6
Central Britt shear zone
Pointe au Baril Station
wR 1801
Bolger Pluton
Parry Sound shear zone
Georgian
Bay
Shawanaga Granodiorite
69
Nobel Granodiorite
Nobel
zU 1346 +69 -39
Parry Sound
0 10 20 30 40 50 km
81°W
80°W

**Figure 19.44.** Generalized geology of the Britt Domain (*modified from* Davidson et al. 1982 and Culshaw et al. 1988, 1991).

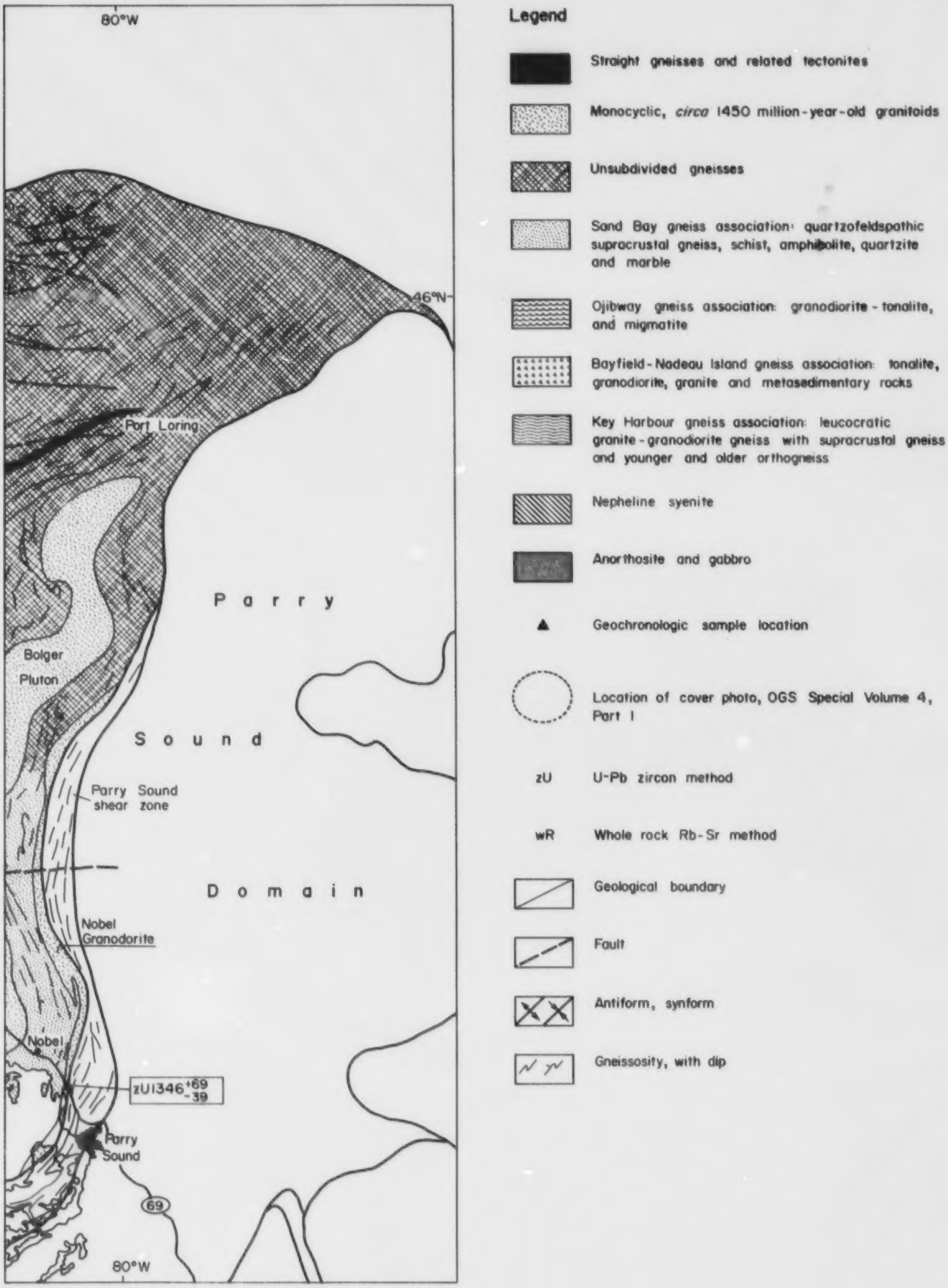

80°W
46°N
Port Loring
Bolger Pluton
Parry Sound Domain
Parry Sound shear zone
Nobel Granodiorite
Nobel
zU1346$^{+69}_{-39}$
Parry Sound
69
80°W
Legend
Straight gneisses and related tectonites
Monocyclic, *circa* 1450 million-year-old granitoids
Unsubdivided gneisses
Sand Bay gneiss association: quartzofeldspathic supracrustal gneiss, schist, amphibolite, quartzite and marble
Ojibway gneiss association: granodiorite-tonalite, and migmatite
Bayfield-Nadeau Island gneiss association: tonalite, granodiorite, granite and metasedimentary rocks
Key Harbour gneiss association: leucocratic granite-granodiorite gneiss with supracrustal gneiss and younger and older orthogneiss
Nepheline syenite
Anorthosite and gabbro
Geochronologic sample location
Location of cover photo, OGS Special Volume 4, Part I
zU U-Pb zircon method
wR Whole rock Rb-Sr method
Geological boundary
Fault
Antiform, synform
Gneissosity, with dip

**Figure 19.44.** Generalized geology of the Britt Domain (*modified from* Davidson et al. 1982 and Culshaw et al. 1988, 1991).

CBSZ show that extensional shear, with top-side-down parallel to the southeast-plunging lineation, took place at high metamorphic grade (Culshaw et al. 1991). Ketchum et al. (1992) have dated, using U-Pb methods on zircon and titanite, pegmatite dikes which cut and are deformed by the extensional fabric in the CBSZ at 988 and 1041 Ma, respectively. These ages constrain the timing of extensional reactivation of the CBSZ.

Figure 19.45 shows an interpretation of the geology between Pointe au Baril and Honey Harbour made by Culshaw et al. (1991) illustrating the relations between the Britt, Go Home and Rosseau domains and the overlying Parry Sound Terrane. All boundaries are shown as thrusts marked by metabasite (including pseudoeclogite)-anorthosite assemblages. The most important thrust is the basal detachment below which lies the parautochthonous "midcontinent" assemblage of polycyclic (greater than 1600 Ma) and monocyclic (ca. 1450 Ma) gneisses (*see* Figure 19.45, unit 1). Above it all, units are probably younger than *circa* 1450 Ma and are metamorphosed only in the Grenvillian event (*see* Figure 19.45, units 2 to 4). Thrust sheets within this duplex-like structure increase in grade and age of metamorphism with structural height, except for the amphibolite-facies sheet (*see* Figure 19.45, unit 5). Culshaw et al. (1991) envisaged the duplex to have been assembled progressively, with the incorporation of the oldest sheets being the most distal. Final crustal thickening above the Britt Domain resulted in heating and subsequent ductile modification of the thrust pile (northwest-folds and extension). Whether the last stage was accompanied by significant displacement along lower thrusts is not known.

### METAMORPHISM

In the Britt Domain, Grenvillian metamorphic overprinting has nearly eradicated granulite-facies assemblages. The latter are primarily preserved as rare orthopyroxene-plagioclase leucosomes in mafic rocks. The largest of the relicts is near Pointe au Baril, where exposures of mafic granulite and corundum-bearing pelitic rocks occur. This is one of several occurrences of high grade assemblages which lie just below the CBSZ (Davidson et al. 1982). Estimates of peak Grenvillian conditions at Key Harbour and Pointe au Baril are 8 to 10 kb at 700° to 800°C (Culshaw et al. 1991), in general agreement with the results of Anovitz and Essene (1990). All estimates on the timing of Grenville metamorphism from the Britt Domain (zircon lower intercepts, monazite and titanite) fall in the interval *circa* 1035 to 1000 Ma (Key Harbour, Nadeau Island and Ojibway polycyclic gneisses and monocyclic plutons; Corrigan et al. 1990; Culshaw et al. 1991). Corrigan (1990) used a lower intercept zircon age of $1015^{+16}_{-17}$ Ma from leucosome in the monocyclic Mann Island Granodiorite to estimate the age of migmatization in the northern Britt Domain. The granulite-facies metamorphism of the Britt Domain is similar in age to the monocyclic granitoids (i.e., 1450 to 1420 Ma; Culshaw et al. 1992; Ketchum et al. 1992).

### MINERAL DEPOSITS

No major mineral occurrences or operating mines have been reported from the Britt Domain. The Bigwood Alkalic Complex is a potential source of nepheline syenite for use in ceramics (Hewitt 1961). The Rutter Monzonite and migmatitic gneisses in the Britt Domain have building stone potential (Lacey 1989; Fouts and Marmont 1989).

## Ahmic Domain

The Ahmic Domain is a small domain sandwiched between the Parry Sound and Kiosk domains. The western half of the domain was mapped in detail by Bright (1990). Bright (1990) recognized an older group of granitoid gneisses and related migmatites and a younger group of monocyclic(?) granodioritic intrusive rocks. The Ahmic Domain is lithologically similar to the Britt Domain which is adjacent to the western margin of the Parry Sound Domain (Davidson et al. 1982, 1985; Bright 1990). The 2 domains may be connected at depth (Davidson et al. 1982; Lindia et al. 1983). Structurally, the Ahmic Domain defines a broad antiform, plunging west beneath the Parry Sound Domain. Rocks of the Ahmic Domain are more highly strained along the contacts with the Parry Sound Domain. No mineral deposits are known in the Ahmic Domain.

## Go Home Domain

### GEOLOGY

Culshaw et al. (1990, 1991) have mapped the western Go Home Domain at 1:50 000 scale. Schwerdtner and Mawer (1982) provide a reconnaissance map of the remainder of the domain.

Culshaw et al. (1990, 1991) recognized 2 gneiss associations within the eastern part of the domain, as described below:

#### Honey Harbour Gneiss Association

The Honey Harbour gneiss association (Figure 19.46) is largely a uniform, granoblastic, migmatitic, pink, granitic or grey granodioritic gneiss (with biotite ± hornblende). These polycyclic orthogneisses commonly have a fabric that reflects the shape of the pluton. Rare pelitic gneiss with sillimanite, garnet and mica, resembles biotite paragneiss of the Key Harbour gneiss association (Britt Domain).

Sheets of metaigneous gabbroic or dioritic, monzonitic and granitic gneiss occur throughout the Honey Harbour gneiss association. They may be monocyclic or polycyclic (Culshaw et al. 1991). Metamorphosed mafic dikes occur sporadically within the Honey Harbour gneiss association as well as common, roughly globular bodies of meta-gabbro, which are restricted to the Honey Harbour gneiss association.

The monocyclic Go Home complex (*see* Figure 19.46) consists mainly of megacrystic (monzo-) granites and granodiorites, containing biotite, hornblende and subordinate garnet. The northwestern boundary of the complex is marked by a thin orange folded sheet of garnet-hornblende monzonite; the centre is underlain by pale grey granodiorite

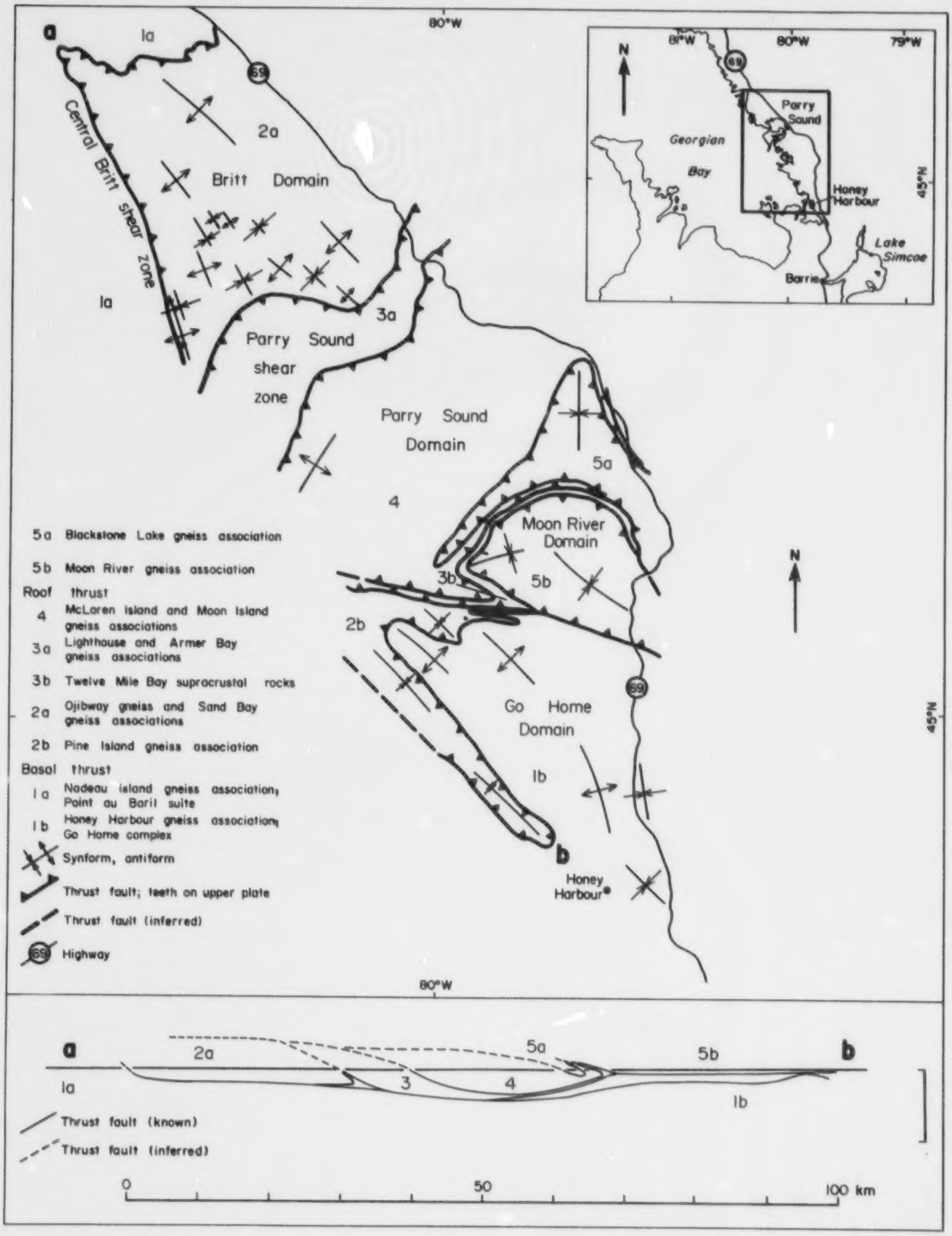

a
Ia
80°W
69
2a
Britt Domain
Central Britt shear zone
Ia
3a
Parry Sound shear zone
Parry Sound Domain
4
5a
Moon River Domain
3b
5b
2b
69
Go Home Domain
Ib
b
Honey Harbour
N
45°N
81°W
80°W
79°W
N
69
Parry Sound
Georgian Bay
Honey Harbour
45°N
Lake Simcoe
Barrie
5a Blackstone Lake gneiss association
5b Moon River gneiss association
Roof thrust
4 McLaren Island and Moon Island gneiss associations
3a Lighthouse and Armer Bay gneiss associations
3b Twelve Mile Bay supracrustal rocks
2a Ojibway gneiss and Sand Bay gneiss associations
2b Pine Island gneiss association
Basal thrust
Ia Nadeau Island gneiss association, Point au Baril suite
Ib Honey Harbour gneiss association, Go Home complex
Synform, antiform
Thrust fault; teeth on upper plate
Thrust fault (inferred)
69 Highway
80°W
a
2a
5a
5b
b
Ia
3
4
Ib
Thrust fault (known)
Thrust fault (inferred)
0
50
100 km

**Figure 19.45.** Map view and structural cross section along Georgian Bay showing relationships between domains and terranes as interpreted by Culshaw et al. (1991). Teeth mark upper plate of thrusts. In cross section, 1 = parautochthonous footwall, 2 to 4 are duplex units, 5 = roof of thrust stack.

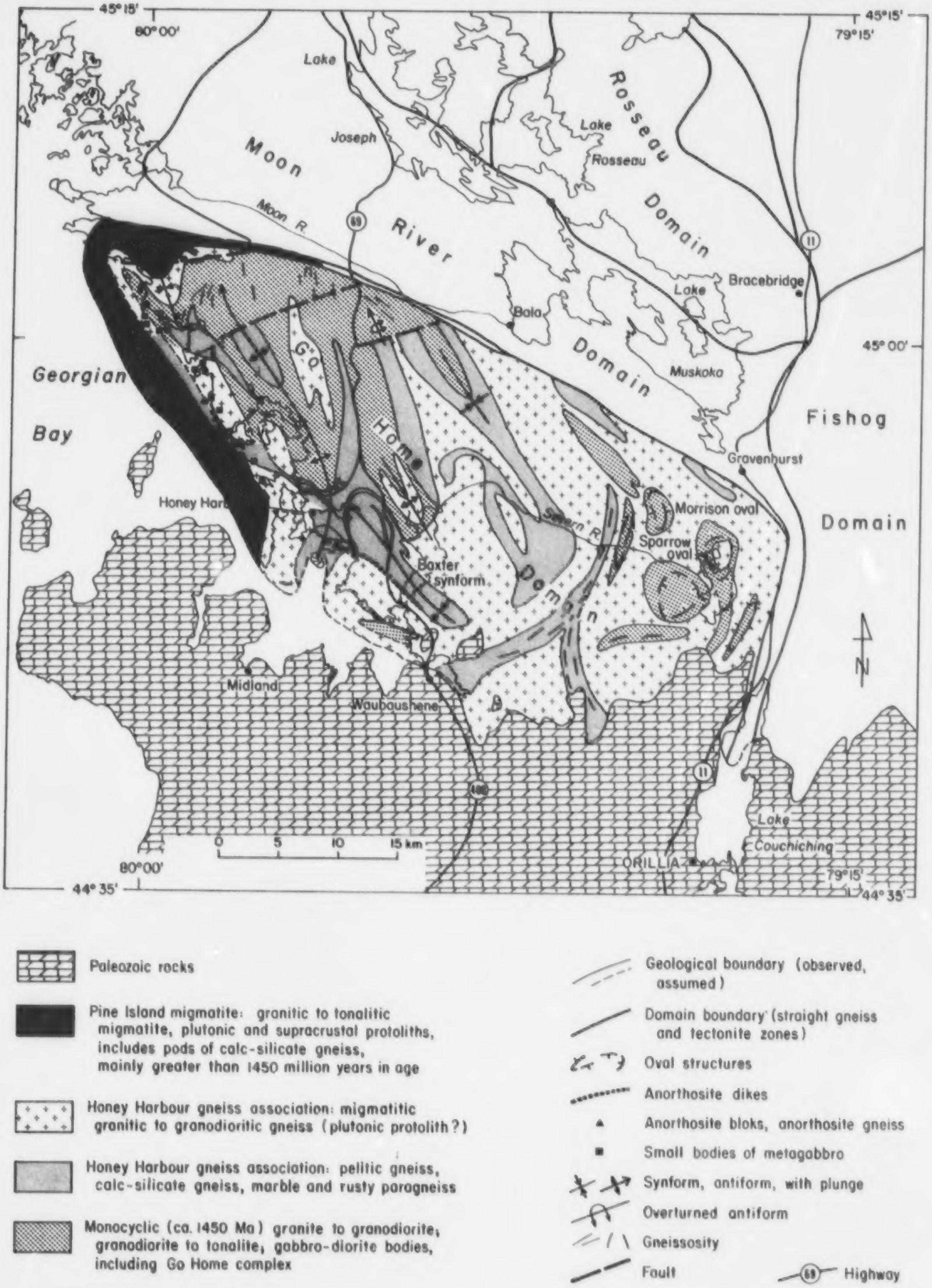

45°15'
80°00'
45°15'
79°15'
Lake
Joseph
Moon River Domain
Moon R.
69
Rosseau Domain
Lake Rosseau
11
Bracebridge
Lake Muskoka
Bala
Domain
45°00'
Georgian Bay
Go Home Domain
Fishog Domain
Gravenhurst
Honey Harbour
Morrison oval
Severn R.
Sparrow oval
Baxter synform
Midland
Waubaushene
400
11
Lake Couchiching
ORILLIA
0 5 10 15 km
80°00'
44°35'
79°15'
44°35'
N
Paleozoic rocks
Pine Island migmatite: granitic to tonalitic migmatite, plutonic and supracrustal protoliths, includes pods of calc-silicate gneiss, mainly greater than 1450 million years in age
Honey Harbour gneiss association: migmatitic granitic to granodioritic gneiss (plutonic protolith?)
Honey Harbour gneiss association: pelitic gneiss, calc-silicate gneiss, marble and rusty paragneiss
Monocyclic (ca. 1450 Ma) granite to granodiorite; granodiorite to tonalite; gabbro-diorite bodies, including Go Home complex
Geological boundary (observed, assumed)
Domain boundary (straight gneiss and tectonite zones)
Oval structures
Anorthosite dikes
Anorthosite bloks, anorthosite gneiss
Small bodies of metagabbro
Synform, antiform, with plunge
Overturned antiform
Gneissosity
Fault
69 Highway

**Figure 19.46.** Generalized geology of the Go Home Domain (*modified from* Culshaw et al. 1991 and Schwerdtner and Mawer 1982).

to tonalite. Elongate bodies of dioritic to gabbroic gneiss lie within the complex. The complex is strongly deformed, having been affected by both regional folds and pervasive ductile deformation. Mafic dikes occur sporadically. These rocks have been tentatively correlated with the *circa* 1420 million-year-old granitoids of the Brandy Lake complex of the Rosseau Domain (Culshaw et al. 1990; *see* Figure 19.46).

#### Pine Island Migmatites

The Pine Island migmatites are characterized by abundant leucosome and are the structurally highest unit in the Go Home Domain. The basal tonalitic migmatitic gneiss resembles the Ojibway gneiss association (Britt Domain). It is overlain by biotite-bearing migmatites of granitic and granodioritic composition and alaskitic pink leucogneiss. There are also thin but widespread pods and layers of calc-silicate and rare leucocratic granitic migmatite with muscovite and sillimanite. The Pine Island migmatites are not cut by the Go Home complex, plutonic members of the Honey Harbour gneiss association, recognizable mafic dikes or gabbro pods. There is no evidence for more than 1 metamorphic event, hence the Pine Island migmatites are likely younger than *circa* 1450 million years in age (Culshaw et al. 1991). The boundary with the Honey Harbour gneiss association is strongly strained and decorated with small bodies of anorthosite, garnet amphibolite and pseudoeclogite. The anorthosites are mostly metre-scale rounded blocks or aggregates of strongly foliated and recrystallized gneiss, with their foliation commonly at a high angle to that in the country rock. Elongate sill-like bodies of anorthosite gneiss also occur (Van Kranendonk 1985, 1987). Culshaw et al. (1991) interpreted the boundary to be a detachment on account of the lack of continuity of units. They also suggested that the upper boundary of the unit may have a similar origin.

### STRUCTURE

Within the Go Home Domain, structures resemble those of the Britt Domain. Northwest-trending folds are coaxial with the stretching lineation. In the Pine Island migmatites at the highest structural level, only Grenvillian structures are present. Below the postulated detachment at their base, earlier deformational events are in evidence where Go Home complex granites cut migmatitic fabrics.

Schwertdner and Mawer (1982) and Schwertdner et al. (1981) identified several oval structures including the Morrison, Matchedash and Sparrow "ovals" (*see* Figure 19.46). These structures are cored by orthogneiss showing relict igneous textures and have subhorizontal gneissosity in their cores.

### METAMORPHISM

The sole estimate of the age of metamorphism is *circa* 1045 Ma (Davidson and van Breemen 1988). However, Culshaw et al. (1991) note that estimates from the Britt Domain fall in the interval *circa* 1035 to 1000 Ma, and the Go Home Domain appears to be similar in this regard.

### MINERAL DEPOSITS

No major mineral deposits are known from the Go Home Domain.

## Rosseau Domain

### GEOLOGY

Rosseau Domain is bounded by a continuous, outward-dipping tectonite zone. Along its northwest margin, it dips beneath the Parry Sound Domain, with a thick, tectonite zone similar to the Parry Sound shear zone being well-developed. The northern part of the domain has been mapped by Macfie and Dixon (1990). The southern part of the domain is only known through the reconnaissance mapping of Davidson et al. (1982), Waddington (1973), Schwerdtner and Waddington (1978) and Schwerdtner and Mawer (1982). The Rosseau Domain is bisected by an east-northeast-trending tectonite zone containing straight gneiss and a band of discontinuous slivers of anorthositic and mafic gneisses up to several kilometres long, yet only a few tens of metres wide (Davidson et al. 1982).

The northwestern part of the Rosseau Domain consists of homogeneous pink, migmatitic gneiss of granitic to granodioritic composition (*see* Figure 19.41). Macfie and Dixon (1990) suggest it is probably of plutonic origin. Numerous, scattered coronitic metagabbro and ultramafic blocks are present in northwestern Rosseau Domain. The only known supracrustal rock from Rosseau Domain is an outcrop of orange marble near the Rosseau–Parry Sound domain boundary (Macfie and Dixon 1990). Metamorphic grade is amphibolite facies.

The southwestern part of the Rosseau Domain consists of a complex of paragneisses and orthogneisses preserved at granulite facies (Davidson et al. 1982). The Lake Muskoka Granite has been dated at 1414 Ma using U-Pb zircon methods (Krogh and Davis 1969). The granite and associated tonalite, monzonite and gabbroic metaplutonic rocks are together termed the Brandy Lake complex (e.g., Schwerdtner and Waddington 1978). Schwerdtner et al. (1977) provide a structural analysis of the Brandy Lake complex.

### MINERAL DEPOSITS

No major mineral deposits are known from this domain, and no mines have operated in the area in the past. Some building stone quarries have operated in the area in the recent past.

## Huntsville Domain

### GEOLOGY

The Huntsville Domain is one of the most extensively mapped domains (Nadeau *in* Davidson et al. 1985; Nadeau 1990; Figure 19.47). The domain consists predominantly of shallowly south-southeast dipping, thin sheets of orthogneiss interleaved with very fine-grained, well-lineated, flaggy, in part layered, high grade gneisses and tectonites (Davidson et al. 1985). The flaggy gneisses are derived from plutonic and sedimentary protoliths.

Most of the orthogneiss in the southern and western parts of the domain (*see* Figure 19.47) is granitic to granodioritic in composition. Locally the orthogneiss grades into a compositionally similar, dark, waxy brown hypersthene gneiss. With increasing deformation, the orthogneiss grades into a fine-grained, equigranular, strongly lineated flaggy tectonite (Davidson et al. 1985).

In the southwestern part of the Huntsville Domain, pink biotite augen granite sheets are juxtaposed with medium grey, recrystallized, homogeneous quartz diorite to granodiorite orthogneisses (*see* Figure 19.47). North of Lake Vernon (*see* Figure 19.47), these metaplutonic rocks enclose a few decametre thick lenses of fine-grained, sugary, pink, leucocratic biotite or garnet-studded monzonitic orthogneiss, which collectively have been termed the Lake Vernon suite (Davidson et al. 1985). Nadeau and van Breemen (1989, 1990) report zircon ages of $1444^{+12}_{-8}$, $1442^{+9}_{-8}$, 1453±6 and $1427^{+16}_{-13}$ Ma from granitic and granodioritic orthogneisses of the Huntsville Domain, confirming the abundance of 1480 to 1420 million-year-old plutons in the Algonquin Terrane. Nadeau and van Breemen (1990) also report a zircon age of $1714^{+123}_{-71}$ Ma (possibly an inherited age) from a quartz monzonite orthogneiss from the Huntsville thrust zone. This age is consistent with Nd-Sm model ages (Dickin and McNutt 1989a) showing the presence of 1.8 billion-year-old crust throughout the Algonquin Terrane.

Although particularly abundant in the southern and western parts of the Huntsville Domain, pelitic and semipelitic gneisses are a widespread but minor unit throughout the domain (*see* Figure 19.47). Davidson et al. (1985) describe them as typically well-foliated, fine- to medium-grained, garnet porphyroblastic gneisses, containing an attenuated planar leucosome and crude layering. Locally associated with these pelitic and semipelitic gneisses are muscovite-biotite-sillimanite quartzofeldspathic gneiss; leucocratic, cream to pink, vaguely layered, quartz-rich feldspathic gneisses; calc-silicate gneisses (diopside-grossularite-scapolite-calcium-plagioclase) enclosing thin marble layers; pink to grey, biotite or hornblende, or both, quartzofeldspathic gneisses; and dark, waxy green, garnet-studded charnockitic gneisses (Davidson et al. 1985), all of probable metasedimentary origin.

The northern part of the Huntsville Domain is characterized by a succession of thin, highly flattened, granulite facies, flaggy tectonites derived from orthogneiss and paragneiss (*see* Figure 19.47). These gneisses enclose discontinuous folded and pulled apart amphibolite and mafic granulite layers, in places likely derived from dikes, and metre- to outcrop-scale lenses which contain coarse-grained igneous textures in their cores. Northeast of Dwight, these metagabbro pods are associated with garnet clinopyroxenite (pseudoeclogite; Davidson et al. 1985). On a mesoscopic scale, this sequence of flaggy tectonites is very heterogeneous, however, at 1:50 000 scale, this sequence has the same lithologic and structural characteristics (Davidson et al. 1985). Rock units form decametre-thick sheets which are difficult to trace along strike and are repeated at different structural levels. Granulite-facies rocks of felsic and intermediate composition predominate over grey hornblende and biotite quartzofeldspathic gneiss. Orthogneisses and paragneisses alternate with gneiss of indecipherable parentage. Centimetre- to metre-scale mafic layers are widespread and are remarkably continuous along strike, even where extremely thin.

## STRUCTURE

Most primary textures have been destroyed, and most rocks show evidence of high shear strain. Lithologic contacts and mineral foliation are conformable. Compositional layering has been tectonically modified, and metaplutonic contacts are also unlikely to be primary (Davidson et al. 1985). Recumbent isoclinal folds and sheath folds are abundant, and combined with high-strain textures, suggests that transposition and tectonic slicing are prevalent. In the western part of the domain, planar fabrics parallel lithologic contacts and dip shallowly to moderately southwesterly; to the east, the strike of the planar fabric swings northeastward. Southeast dips become shallower to the south with increasing distance from the Huntsville–Novar domain boundary. Lineations are variably oriented (Davidson et al. 1985).

Davidson et al. (1985) also report late, metre-wide extensional shears in the domain (and in the southern Algonquin Terrane). These shears disturb pre-existing structures. Where these shears transect granulite-facies rocks, they are associated with metre-scale zones of retrogression to amphibolite facies. The shears are preferentially developed in well-layered rocks. Several subparallel shears are commonly developed in large exposures. East-northeast-striking and steeply south-dipping shears are dominant in the Huntsville Domain. Late granitic pegmatites are commonly emplaced along the shear plane. Nadeau and van Breemen (1989, 1990) report a zircon date from a pegmatite within one of these shear zones of $1039^{+4}_{-2}$ Ma and noted that this demonstrates that the initiation of uplift and associated extension was coincident with the late stages of regional northwest-directed thrusting.

## METAMORPHISM

The presence of highly attenuated granitoid stringers in amphibolite-facies rocks suggests high-grade metamorphism occurred prior to flattening, and granulite-facies rocks may also record an earlier metamorphism (Davidson et al. 1985). New zircon growth occurred at $1030^{+50}_{-20}$ Ma in a granulite-facies orthogneiss located in the Huntsville Domain (van Breemen et al. 1986) and was interpreted by them as dating the granulite-facies metamorphism. Nadeau and van Breemen (1990) note that high metamorphic temperatures persist for at least 35 million years after peak metamorphism, hence granulite-facies metamorphism may have been initiated as long ago as 1065 Ma.

## MINERAL DEPOSITS

As is typical of the domains present within the Algonquin Terrane, few mineral deposits have been noted in the Huntsville Domain. Several flagstone quarries are present within the tectonized northern part of the domain (Fouts and Marmont 1989), and potential for further sites in the flaggy gneisses of the northern part of the domain (*see* Figure 19.47) is high.

## Kiosk Domain

### GEOLOGY

Structural trends in this domain are predominantly east-northeasterly; gneissosity dips moderately southeast, and strongly developed, penetrative mineral lineation plunges downdip (Davidson and Morgan 1981). As originally defined by Davidson and Morgan (1981), the Kiosk Domain included the weakly structured Powassan Monzonite, however Davidson and Grant (1986) redefined the northwestern boundary of the domain to exclude the Powassan Monzonite.

This domain is, for the most part, poorly exposed, but its structural pattern is well displayed by the pattern of aeromagnetic anomalies (Davidson and Morgan 1981). The metaplutonic rocks include a cream to pale, salmon pink hornblende metamonzonite to metamorphosed quartz monzonite with conspicuous, wine-red garnet porphyroblasts. More deformed equivalents of this rock become yellow-green in colour and locally contain hypersthene (Davidson and Morgan 1981), however, the intraplutonic gneisses show no evidence of being at granulite facies (Davidson and Morgan 1981). The southern boundary of the Kiosk Domain appears to be one of change to a less regular structural trend and to the higher metamorphic grade of the Huntsville, McCraney and McClintock domains (Davidson and Morgan 1981).

### MINERAL DEPOSITS

The only operating mine in the Central Gneiss Belt is the Cal Graphite graphite mine in Butt Township, east of Highway 11, near the boundary of Algonquin Provincial Park (Garland 1991). The mine started operating in 1990 and is producing 2 800 tonnes per day of coarse flake graphite (Russell 1991). Another graphite prospect lies in Ryerson Township, near the eastern boundary of the Parry Sound Domain (Garland 1987). The abundance of graphite occurrences in the Kiosk Domain may reflect the abundance of granulite-facies metasedimentary gneisses. Potential for other prospects in the domain remain high. Two flagstone quarries are also present in the northern Kiosk Domain along the boundary with the Powassan Domain (Fouts and Marmont 1989).

## McClintock Domain

### GEOLOGY

Culshaw et al. (1983) report that the domain is underlain, in large part, by enderbitic and charnockitic orthogneiss. Metasedimentary gneisses in granulite facies form narrow, sinous horizons between the orthogneiss bodies and display northwest-verging recumbent folding (Culshaw et al. 1983). Calc-silicate gneiss and marble are present locally interlayered with pelitic gneisses. Nadeau and van Breemen (1989) report a zircon age of 1027±5 Ma from a late syntectonic pegmatite from the McClintock–Huntsville domain boundary.

### MINERAL DEPOSITS

A gabbroic intrusion in McClintock Township with ultramafic phases has been examined recently for its nickel and platinum group element potential (Garland et al. 1989, 1990). It is part of a group of *circa* 1170 million-year-old gabbroic intrusions found throughout the Algonquin Terrane (*see* Figure 19.20d). A flagstone quarry is operating along the Kawagama shear zone, which marks the southern boundary of the domain with the Fishog Domain (Fouts and Marmont 1989), and another quarry is operating along the McCraney–McClintock boundary (Fouts and Marmont 1989). Most of the northern and eastern portions of the McClintock Domain lie within Algonquin Provincial Park.

## McCraney Domain

### GEOLOGY

An east- to northeast-trending belt of flaggy, predominantly quartzofeldspathic tectonites separates the McCraney Domain from the Novar and Huntsville domains to the east. Internally, pink to pale green, fine- to very fine-grained, quartzofeldspathic gneisses are abundant, particularly, in the western part of the domain (Culshaw et al. 1983). Some grade into granitoid orthogneiss with potassium feldspar augen, but others are interlayered with biotite-hypersthene gneiss and pelitic gneiss and contain greater than 50% quartz, suggesting a metasedimentary origin (Culshaw et al. 1983). Mafic rocks are subordinate. Much of the terrane is at granulite facies. The internal structure of the domain is characterized by intermediate- to large-scale folds, coaxial with a dominant east-southeast lineation (Culshaw et al. 1983).

### MINERAL DEPOSITS

No major mineral deposits are known from the McCraney Domain. Several flagstone quarries are present in the southern part of the domain just west of the border with Algonquin Provincial Park (Fouts and Marmont 1989). Most of the domain lies within Algonquin Provincial Park.

## Novar Domain

### GEOLOGY

Metaplutonic rocks, separated by septa of supracrustal gneisses, underlie a large part of the Novar Domain (Culshaw et al. 1983; *see* Figure 19.47). The largest orthogneiss body, near Novar, is composed mainly of homogeneous, hornblende-rich garnetiferous metamorphosed quartz diorite, which locally grades into mafic dioritic to gabbroic phases. The dioritic rocks are fine- to medium- grained, well-foliated, and although recrystallized, locally retain relict igneous textures (Davidson et al. 1985). A weakly migmatitic, pink, granitic augen gneiss is locally associated with the metamorphosed quartz dioritic rocks. Davidson et al. (1985) report that charnockitic orthogneiss is present sporadically throughout the Novar Domain, and forms kilometre-size masses enclosed in commonly migmatitic granitoid or quartzofeldspathic gneiss. Contacts with surrounding rocks are gradational, and the charnockitic gneisses are heterogeneous. The largest mass is north of Lake Vernon (Davidson et al. 1985). Davidson et al.

(1985) also report several exposures of anorthositic gneiss in the northwest part of Lake Vernon. These occurrences define a seam-like body that can be traced more than 7 km along strike, and which connects with a larger body of garnet amphibolite, metagabbro and garnet clinopyroxenite that encloses a few, metre-thick, discontinuous meta-anorthosite layers (Davidson et al. 1985). Deeply weathered ultramafic rock occurs in contact with well-foliated, homogeneous garnet amphibolite (metagabbro?) (Davidson et al. 1985). Small bodies of well-foliated granitic and granodioritic orthogneiss are widespread within the Novar Domain.

Novar Domain metasedimentary gneisses consist of quartz-rich feldspathic (arkose?), semipelitic and lesser pelitic varieties (Davidson et al. 1985). Calc-silicate gneiss and marble have not been reported from the Novar Domain. A nondescript quartzofeldspathic biotite gneiss, with or without garnet, is a common rock type in the Novar Domain, and may also be supracrustal in origin.

## STRUCTURE

The Novar Domain differs in both style and orientation from that of the Huntsville and Seguin domains. North-northwest

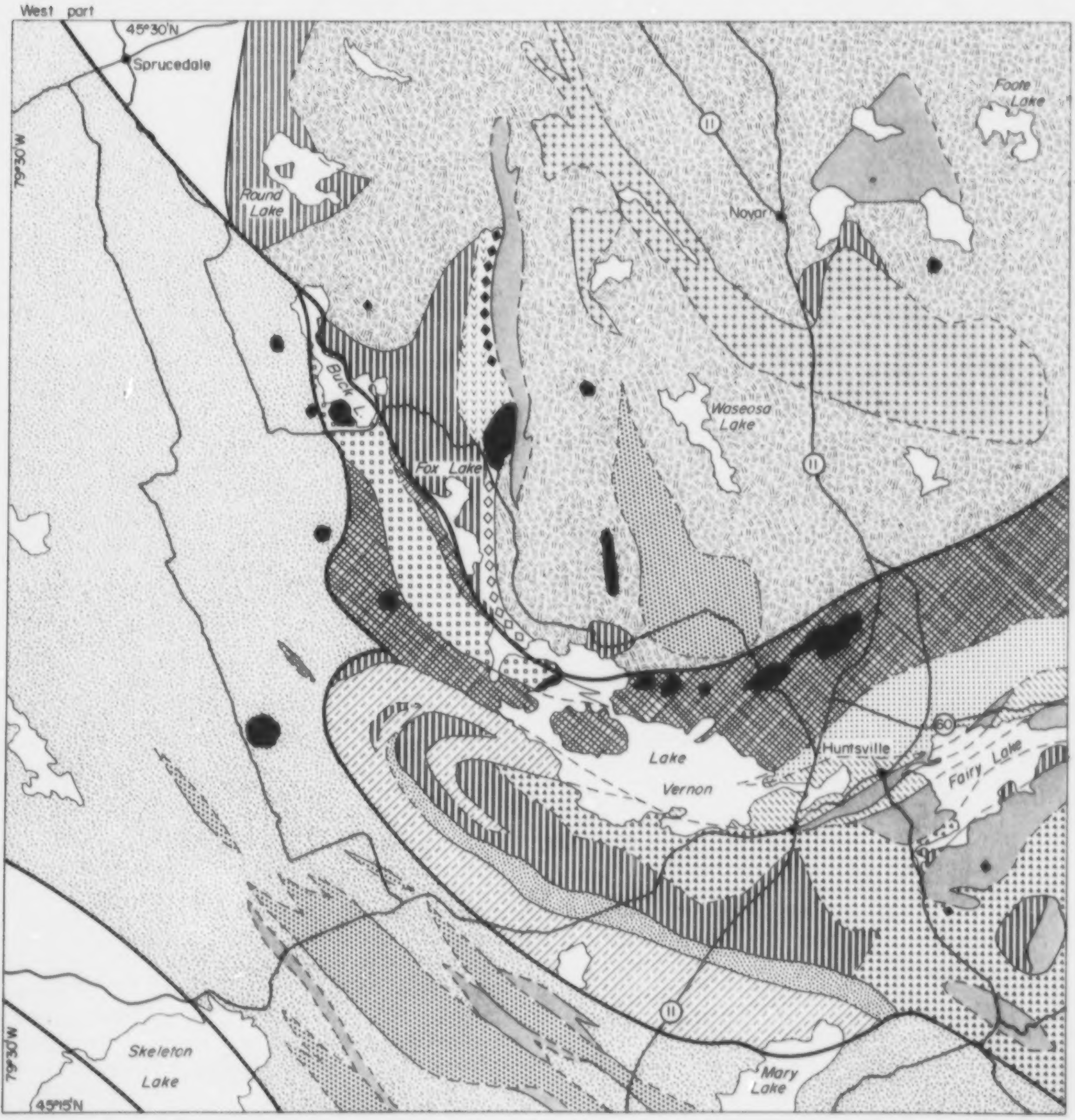

West part
45°30'N
Sprucedale
79°30'W
Round Lake
Buck L.
Fox Lake
11
Novar
Foote Lake
Waseosa Lake
11
Lake Vernon
Huntsville
60
Fairy Lake
11
Skeleton Lake
Mary Lake
79°30'W
45°15'N

**Figure 19.47.** Generalized geology of the Huntsville Domain (*modified from* Davidson et al. 1985), showing complex relations typical of Algonquin Terrane domains.

trends prevail throughout much of the southern half of the Novar Domain but swing to the northeast along the western edge (Davidson et al. 1985). Dips vary from moderate to vertical. Lineation attitudes vary widely. As a whole, the Novar Domain structure is highly discordant compared to other domains, a feature which Davidson et al. (1985) suggested was inherited from an earlier phase of deformation.

### MINERAL DEPOSITS

No major mineral deposits have been reported from the Novar Domain.

## Opeongo and Powassan Domains

### GEOLOGY

These 2 domains were mapped in a reconnaissance fashion by Lumbers (1976a, 1976b, 1982). Davidson and Grant (1986) mapped the western part of the Opeongo Domain and the southern Powassan Domain. They included the southern Powassan Domain within the Kiosk Domain, although they showed a domain boundary between the Kiosk Domain and an unnamed domain to the north in their Figure 86.1, coincident with the Powassan–Kiosk boundary used in this report (*see* Figure 19.2). The 2 domains are discussed together as

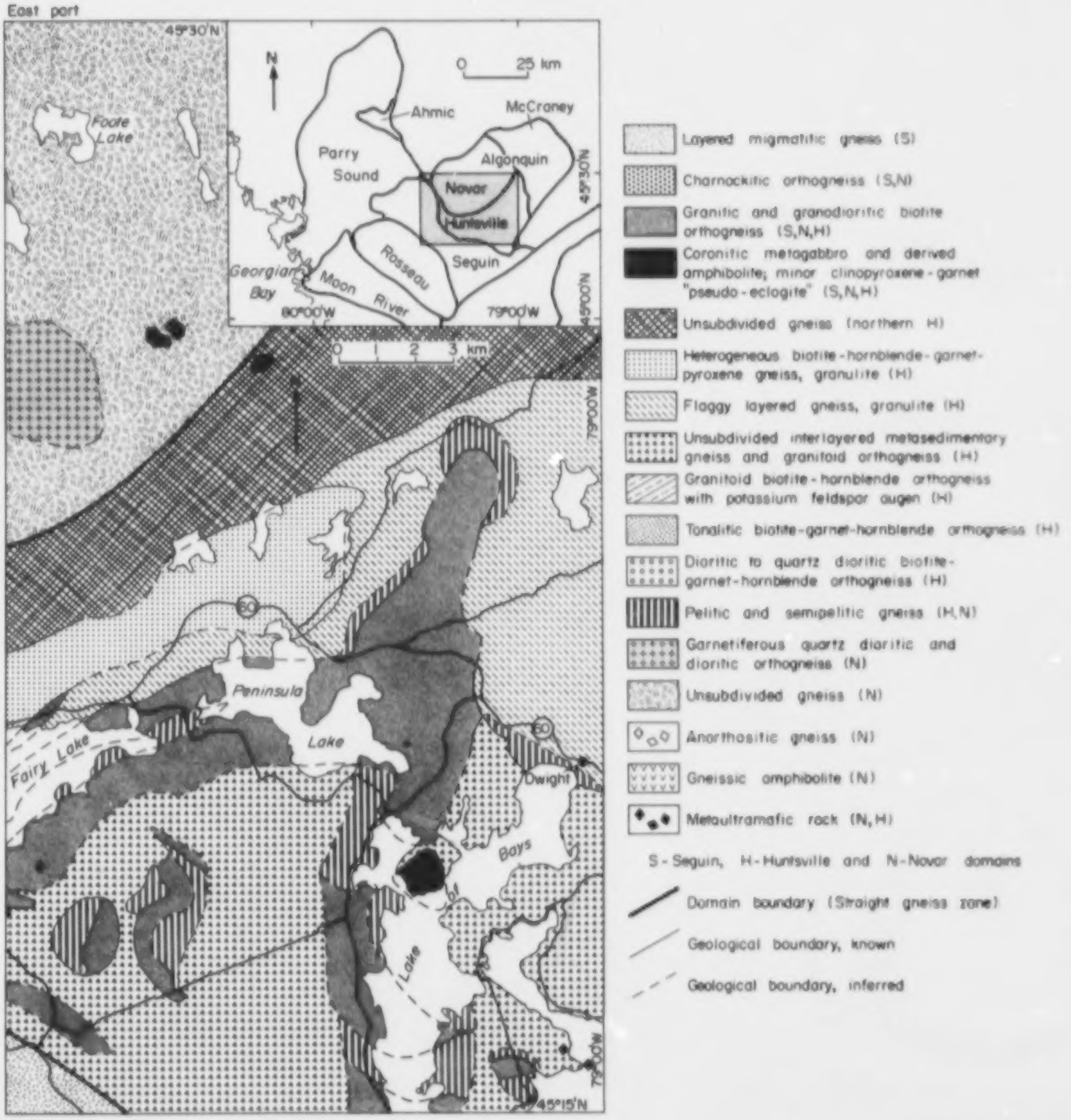

East port
45°30'N
Foote Lake
N
0 25 km
Ahmic
McCraney
Parry Sound
Algonquin
Novar
Huntsville
Rosseau
Seguin
Georgian Bay
Moon River
80°00'W
79°00'W
45°30'N
45°00'N
0 1 2 3 km
79°00'W
60
Peninsula
Fairy Lake
Lake
60
Dwight
Bays
of
Lake
79°00'W
45°15'N
Layered migmatitic gneiss (S)
Charnockitic orthogneiss (S,N)
Granitic and granodioritic biotite orthogneiss (S,N,H)
Coronitic metagabbro and derived amphibolite; minor clinopyroxene-garnet "pseudo-eclogite" (S,N,H)
Unsubdivided gneiss (northern H)
Heterogeneous biotite-hornblende-garnet-pyroxene gneiss, granulite (H)
Flaggy layered gneiss, granulite (H)
Unsubdivided interlayered metasedimentary gneiss and granitoid orthogneiss (H)
Granitoid biotite-hornblende orthogneiss with potassium feldspar augen (H)
Tonalitic biotite-garnet-hornblende orthogneiss (H)
Dioritic to quartz dioritic biotite-garnet-hornblende orthogneiss (H)
Pelitic and semipelitic gneiss (H,N)
Garnetiferous quartz dioritic and dioritic orthogneiss (N)
Unsubdivided gneiss (N)
Anorthositic gneiss (N)
Gneissic amphibolite (N)
Metaultramafic rock (N,H)
S-Seguin, H-Huntsville and N-Novar domains
Domain boundary (Straight gneiss zone)
Geological boundary, known
Geological boundary, inferred

**Figure 19.47.** Generalized geology of the Huntsville Domain (*modified from* Davidson et al. 1985), showing complex relations typical of Algonquin Terrane domains.

the rock types are similar, and the mutual boundaries are uncertain.

Both domains are underlain by large, *circa* 1450 million-year-old monzonitic plutons, including the Powassan, Bonfield, Opeongo Lake and Wilkes monzonites, as well as the "Algonquin Batholith" which underlies most of the eastern and northern Opeongo Domain. Between these plutons are septa of amphibolite-facies paragneisses, including local bands of calc-silicate rocks and marbles. Some of the latter near the Mattawa River contain brucite (Harding 1946).

Aluminous and calcic paragneisses are easily identified by virtue of their distinctive mineral assemblages. The former contain abundant graphite, pyrite and quartz, 2 feldspars, violet garnets and biotite, locally, with sillimanite or green spinel. Metaquartzite locally occurs in the eastern Powassan Domain. The calc-silicate gneisses contain quartz, plagioclase, scapolite, grossularite and diopside assemblages. Marble bands contain graphite and locally brucite or wollastonite. Most of these paragneisses occur as thin layers within nondescript quartzofeldspathic gneisses. The latter may also be metasedimentary in origin.

Monzonite suite plutons such as the Powassan Monzonite are compositionally heterogeneous and are composed of garnet-hornblende metamonzonite and migmatitic metamorphosed quartz monzonite with minor metatonalite and metadiorite (locally garnetiferous). Descriptions and chemical data on the Powassan, Bonfield and Wilkes monzonites are given in Lumbers (1971a). Monzonitic plutons are more abundant in the northern Powassan and eastern Opeongo domains, along the Ottawa–Bonnechere Graben. This abundance may simply reflect a slightly deeper exposure level, perhaps related to movement along the graben-bounding faults.

The Eau Claire meta-anorthosite (Calvin body *of* Lumbers 1976c) in the Powassan Domain, near Mattawa, is slightly older than the Rankin Monzonite and is the largest anorthosite body present in either domain. Lumbers (1976c) reports nepheline syenite in association with the anorthosite, and metasomatism of the anorthosite by the nepheline syenites. The syenites may be related to the *circa* 1290 to 1250 million-year-old nepheline syenite suite of the Central Metasedimentary Belt (Lumbers 1976c; Lumbers et al. 1990). Two small anorthosite bodies occur in Head Township in the Powassan Domain. The Eau Claire meta-anorthosite has a sill-like form and is somewhat similar to the Mercer and St. Charles bodies in the Nepewassi Domain. A small anorthositic body about 5 km east of the Eau Claire body is being explored for vanadium-iron-titanium mineralization (Garland et al. 1989).

### STRUCTURE

The plutons form elongate domal masses, with the metasedimentary and other gneisses wrapping around the masses. Well-developed structural trends in the gneisses are only found near the Parry Sound and Kiosk domain boundaries. The northern part of the Powassan Domain, and much of the central Opeongo Domain, are cut by east-trending faults of the Ottawa-Bonnechere graben system and the related Grenville Swarm diabase dikes.

### METAMORPHISM

Lumbers (1982) reports that most of the area underlain by the Opeongo Domain is preserved at upper amphibolite facies, with possible granulite-facies rocks being present near the Ottawa River in Ross Township.

### MINERAL DEPOSITS

Like the Kiosk Domain, graphite prospects occur throughout the Powassan Domain and are related to the abundant metasedimentary gneisses. The Laurier Township prospect (Garland 1987) lies in a septum between 2 lobes of the Powassan Monzonite. The Maria Township prospect (Garland 1987; Russell 1991) lies near the Ottawa River, in metasedimentary gneiss west of the Rankin Monzonite. Much of the northern and eastern Powassan Domain is covered by drift, and further occurrences may be present. Feldspar and mica have been mined from the Powassan Domain near Mattawa (Harding 1946). Other industrial mineral potential includes brucite and wollastonite. Much of the western Opeongo and southeastern Powassan domains lie within Algonquin Provincial Park.

## Fishog Domain

### GEOLOGY

The Fishog Domain was first defined by Easton (1983, 1986b, 1990a) who noted that lithologically it differed from "Muskoka Domain" rock types farther to the east, and who noted its distinctive aeromagnetic character. The southeastern Fishog Domain has been mapped by Easton (1986b, 1986c, 1987d, 1988g, 1990a). Schwerdtner and Mawer (1982) mapped the western part of the domain in reconnaissance fashion. To the north, the domain is ill-defined. South of Dorset, it is wedged between the Kawagama shear zone and the northern limit of the Central Metasedimentary Belt Boundary Zone, and the boundary between the southern McClintock and northern Fishog domains cannot be easily defined.

The oldest rocks in the domain are migmatitic tonalite gneisses of unknown protolith, and a sequence of metasedimentary gneisses that contain quartz-rich and calc-silicate-rich layers. These are intruded by a suite of monzonitic plutons, which are locally cut by thin anorthosite sills, such as the Head River meta-anorthosite (Easton 1983, 1990b; Van Kranendonk 1985, 1987). A suite of younger, granodiorite and magnetite-bearing, syenogranite plutons cut all other units in the area. These syenogranite bodies can be traced west to the Kashe shear zone, which marks the boundary with the Go Home Domain. Most of the plutonic bodies are migmatitic. Chemical analyses of plutons from the Fishog Domain are given in Easton (1987d, 1990a).

No geochronology is available from the Fishog Domain. If the monzonite plutons are similar to the *circa* 1450 million-year-old monzonites found throughout the Algonquin Terrane, then the syenogranites could be manifestations of the *circa* 1350 to 1320 million-year-old magmatic episode. Overall, the rocks of the Fishog Domain resemble the Pine Island migmatites described by Culshaw et al. (1991) from the Go Home Domain, which they noted are probably younger than *circa* 1450 million years in age.

### STRUCTURE

In the central and western parts of the domain, the plutons form elongate domes with relatively little-deformed cores and attenuated margins. Metasedimentary and migmatitic tonalite gneisses form septa between the larger plutons. To the east, structures become more attenuated and are rotated into parallelism with the Central Metasedimentary Belt Boundary Zone (CMBBZ). In addition, a prominent southeast-plunging lineation becomes more common, and more intense, near the CMBBZ. In the western CMBBZ, it is apparent that some of the tectonites are derived from deformed equivalents of Fishog Domain rocks (Easton 1986b, 1986c, 1987d, 1988g, 1990a). However, the source of many of the tectonites in the CMBBZ cannot be identified.

### METAMORPHISM

Much of the eastern Fishog Domain is in upper amphibolite facies, although some retrograded granulite-facies rocks may be present in Anson and Longford townships (Easton 1987d). Metamorphic grade of the western part of the domain is not known. Certainly in the eastern Fishog Domain, it cannot be determined if the upper amphibolite-facies conditions we see reflect late-tectonic retrogression and fluid movement along the CMBBZ, or simply the absence of granulite-facies metamorphism in the domain. If the latter is correct, this would make the Fishog Domain distinctly different from the bulk of the Algonquin Terrane.

### MINERAL DEPOSITS

Easton (1986b, 1986c, 1987d, 1988g, 1990a) summarizes the mineral potential of the eastern Fishog Domain. No major mineral deposits have been found to date, although there is some potential for gold in the migmatitic tonalite gneisses, as well as potential for building stone in the more accessible parts of the domain.

# PARRY SOUND TERRANE

## Parry Sound Domain

### INTRODUCTION

The Parry Sound Domain is unique in the Central Gneiss Belt in containing large volumes of mafic rock, anorthosite and marble, and in being metamorphosed, relatively uniformly, to granulite facies. It is also the most extensively mapped part of the Central Gneiss Belt, with recent detailed mapping studies by the Ontario Geological Survey (McRoberts et al. 1988; Easton 1989d; Bright 1990; McRoberts 1991) and the Geological Survey of Canada (Culshaw et al. 1989, 1990) covering much of the domain at 1:50 000 scale or better. Both the Ontario Geological Survey studies in the interior of the Parry Sound Domain and the Geological Survey of Canada studies along the coastline of Georgian Bay suggest a broad-scale subdivision of the Parry Sound Domain into a dominantly mafic western part and a dioritic eastern part, separated, at least in the north, by a band of marble tectonic breccia (Figure 19.48).

### GEOLOGY

The contact of the Parry Sound Domain and the underlying Britt Domain (Algonquin Terrane) is a wide zone of tectonites (*see* Figure 19.24), termed the Parry Sound shear zone (PSSZ), derived mainly from rocks of the Parry Sound Domain. As noted by Culshaw et al. (1989), anorthositic masses in the Parry Sound Domain are concentrated along the upper margin of the PSSZ and may have been incorporated into the Parry Sound Domain during thrusting. Tectonite zones are locally present along the lithologically gradational contact with the overlying Seguin and Moon River domains but are less extensive than in the PSSZ. Most lithologic contacts are tectonic, but where deformation is less severe, relict original contacts and textures are present (*see* Figures 19.35, 19.36 and 19.37). Any model for the formation of the Parry Sound Domain must account for: 1) the "stratigraphy" across the Parry Sound Domain; 2) crosscutting relationships indicating a history of mafic magmatic activity; 3) abundance of high strontium marbles; 4) the presence of similar rocks in the Moon River and Sequin domains; and 5) the Nd-Sm data of Dickin and McNutt (1989a, 1989b), indicating *circa* 1.4 billion-year-old crustal extraction of the Parry Sound Domain, Moon River and Seguin rocks compared to 1.8 Ga for surrounding domains.

### GNEISS ASSOCIATIONS

#### Lighthouse and Armer Bay Gneiss Associations

Along Georgian Bay in the westernmost Parry Sound Domain, Culshaw et al. (1989, 1990, 1991) have recognized an amphibolite-facies assemblage of interlayered metabasites (possibly metavolcanic rocks), metaquartzite, marble, pelite, miscellaneous quartzofeldspathic gneiss and a tonalitic-granodioritic orthogneiss. The metaquartzite and orthogneiss are cut by mafic dikes. On Parry Island, there are 2 roughly equivalent packages of this type lying within the PSSZ, the Lighthouse and Armer Bay gneiss associations of Culshaw et al. (1989, 1990, 1991; *see* Figure 19.48). Culshaw et al. (1991) interpreted the boundary between the Parry Sound and Britt domains to be a thrust because the metaquartzite, the tonalitic-granodioritic orthogneiss and the anorthosite contain crosscutting mafic dikes, whereas the underlying Ojibway and Sand Bay gneiss associations of the Britt Domain do not.

A prominent belt of mylonites separates the Lighthouse and Armer Bay gneiss associations from the overlying granulite-facies McLaren Island gneiss association (*see* Figure 19.48). The mylonites are a continuation of those exposed at the town of Parry Sound (Davidson et al. 1982). Culshaw et al. (1991) interpret this mylonite belt as a thrust, since the belt is a lithologic break and marks an apparent metamorphic inversion.

To the north, these 2 gneiss associations have not been identified, although Bright (1990) reports metaquartzite

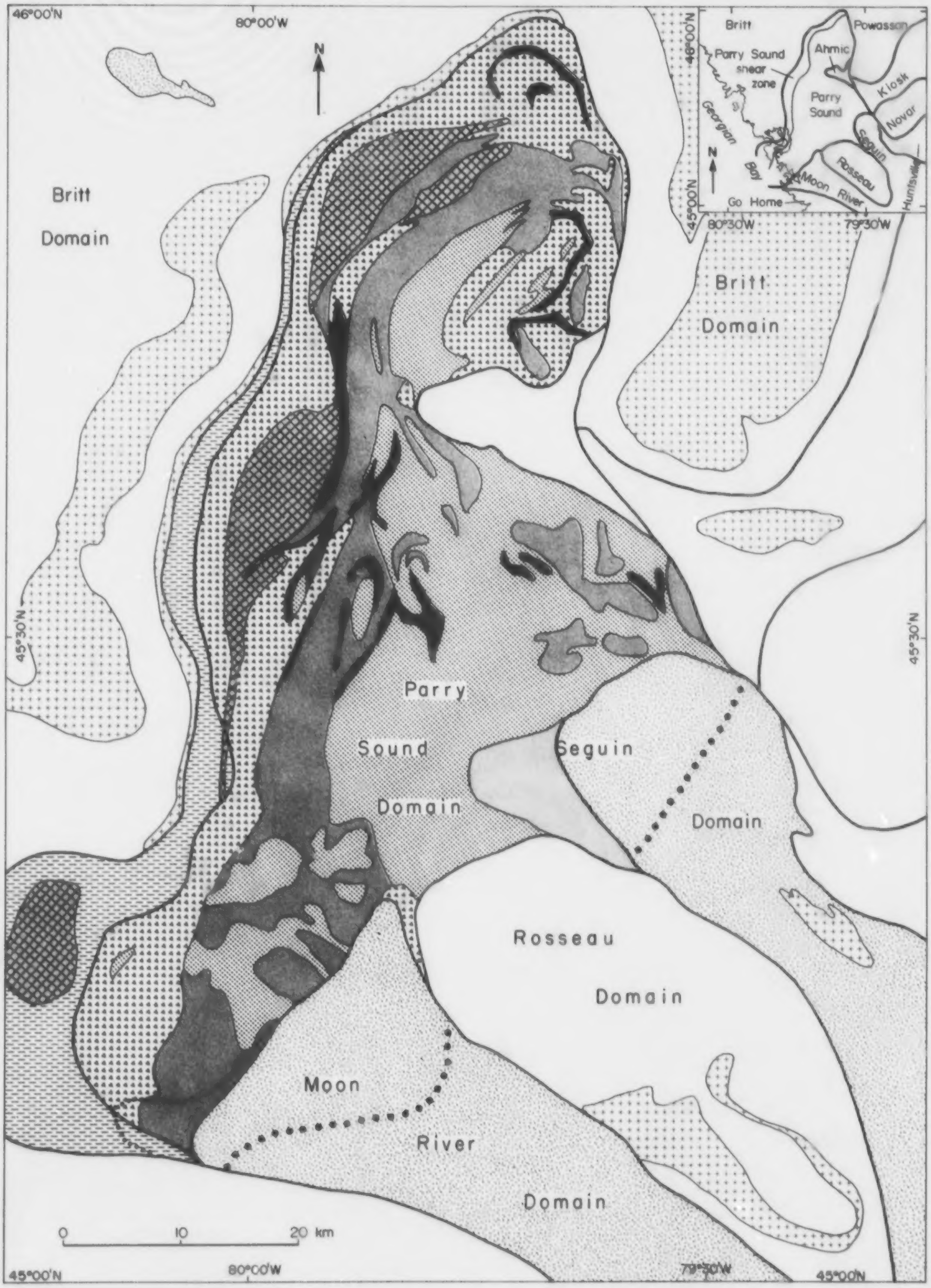

46°00'N
80°00'W
N
Britt
Domain
Britt
Domain
Parry
Sound
Domain
Seguin
Domain
Rosseau
Domain
Moon
River
Domain
45°30'N
45°30'N
0
10
20 km
45°00'N
80°00'W
79°30'W
45°00'N
Britt
Powassan
Parry Sound
shear
zone
Ahmic
Parry
Sound
Kiosk
Novar
Georgian
Bay
Seguin
Rosseau
Moon
River
Huntsville
Go Home
N
46°00'N
45°00'N
80°30'W
79°30'W

**Figure 19.48.** Generalized geology of the Parry Sound Domain based on mapping of Bright (1990), Culshaw et al. (1989, 1990, 1991), McRoberts (1991), McRoberts and Tremblay (1991) and Easton (unpublished data).

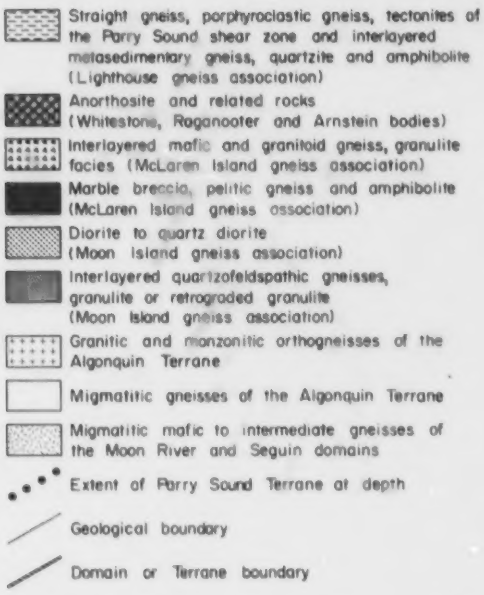

Straight gneiss, porphyroclastic gneiss, tectonites of the Parry Sound shear zone and interlayered metasedimentary gneiss, quartzite and amphibolite (Lighthouse gneiss association)
Anorthosite and related rocks (Whitestone, Raganooter and Arnstein bodies)
Interlayered mafic and granitoid gneiss, granulite facies (McLaren Island gneiss association)
Marble breccia, pelitic gneiss and amphibolite (McLaren Island gneiss association)
Diorite to quartz diorite (Moon Island gneiss association)
Interlayered quartzofeldspathic gneisses, granulite or retrograded granulite (Moon Island gneiss association)
Granitic and monzonitic orthogneisses of the Algonquin Terrane
Migmatitic gneisses of the Algonquin Terrane
Migmatitic mafic to intermediate gneisses of the Moon River and Seguin domains
Extent of Parry Sound Terrane at depth
Geological boundary
Domain or Terrane boundary

**Figure 19.48.** Generalized geology of the Parry Sound Domain based on mapping of Bright (1990), Culshaw et al. (1989, 1990, 1991), McRoberts (1991), McRoberts and Tremblay (1991) and Easton (unpublished data).

layers within the Parry Sound shear zone in the western part of the Whitestone Lake area, and quartzite layers are locally present in the Parry Sound shear zone in the Ferrie area (McRoberts and Tremblay 1987, 1991). Poor exposure, combined with strong tectonic overprinting in the PSSZ, makes distinguishing the Lighthouse and Armer Bay gneiss associations from the McLaren Island gneiss association difficult in the interior.

### McLaren Island Gneiss Association

The McLaren Island gneiss association (Culshaw et al. 1989, 1990, 1991) is formed of granitoid and mafic units interlayered at map to outcrop scale. The mafic units include dikes and units of probable supracrustal and plutonic origin. This belt also includes horizons of retrograde- and granulite-facies mylonitic gneiss and a thin band of metapelite. This unit has been mapped and described by McRoberts et al. (1988) and McRoberts (1991), who have recognized layered gabbroic bodies within the sequence, containing minor ultramafic and anorthositic phases. Sillimanite-bearing pelitic gneisses also occur locally within these sequence, as do a variety of deformed, white to pink granitoid gneisses. Geochemical data on these rocks is provided in McRoberts and Tremblay (1991) and McRoberts (1991). Easton (1989d, 1989b) noted that 2 types of gneisses are present: a tholeiite suite, possibly MORB, showing less than 51% $SiO_2$, greater than 1.2% $TiO_2$ and less than 16% $Al_2O_3$; and an island-arc tholeiite suite of generally basaltic andesite composition with 53% $SiO_2$, less than 1.2% $TiO_2$ and low chromium and nickel values. The rare earth geochemistry of these suites is also distinctive (Figure 19.49; Easton, unpublished data).

### Whitestone Lake Structural Complex

Bright (1990) defined a marble breccia and tectonite zone in the central Parry Sound Domain, which he likened to the Denna Lake Structural Complex of Easton (1990a) in the CMBBZ. This marble breccia zone can be traced north into the Ferrie area (McRoberts and Tremblay 1987, 1991), where it thins and becomes discontinuous, and south into the McKellar (Lacy 1960; Easton, unpublished data) and Manitouwabing areas (McRoberts et al. 1988; McRoberts 1991). Marbles within the zone have high Sr contents, 800 to 1200 ppm in contrast to lower values of 300 to 400 ppm for Central Metasedimentary Belt marbles (Easton, unpublished data). The Whitestone Lake Structural Complex separates the lithologically distinct McLaren and Moon Island gneiss associations and may be a thrust.

### Moon Island Gneiss Association

Southward and eastward, the Parry Sound Domain is dominated by metaplutonic rocks of the Moon Island gneiss association (*see* Figure 19.48). These have more intermediate compositions and are more abundant than granitoid rocks in the McLaren Island gneiss association to the north and contain a unique set of plagioclase-rich mafic dikes (Culshaw et al. 1991). Diorite and quartz diorite plutons, for example, the McKellar Diorite (Dallmeyer and Sutter 1980; Connare and McNutt 1985; Connare 1986; McRoberts 1991), make up the bulk of this terrane. In the Manitouwabing area, deformation is restricted to the margins of the plutons with locally developed intrusion breccias (*see* Figure 19.37; McRoberts et al. 1988; McRoberts 1991). Geochemical data on these rocks is provided *in* McRoberts and Tremblay (1991), McRoberts et al. (1988) and Connare (1986). McRoberts (Ontario Geological Survey, personal communication, 1989) and Easton (1989d, 1989b) distinguished the diorite and leucocratic gneisses of the McKellar suite geochemically. Although both are similar in major element chemistry and show calc-alkalic affinities, they differ considerably in their trace element and rare earth element geochemistry (*see* Figure 19.49). van Breemen et al. (1986) report a date of 1425±75 Ma for the McKellar Diorite.

### Retrograded Gneiss Association

Along Georgian Bay, Culshaw et al. (1991) noted that a zone of Parry Sound Domain gneisses, reworked to amphibolite facies, abuts the Go Home and Moon River domains. Between these and the anorthosite, which lies along the boundary, is a single supracrustal unit (*see* Figure 19.48) composed of metaquartzite, amphibolite and paragneiss similar to, but much thinner than, the Lighthouse gneiss association. Altogether, these rocks, and the similarity of rocks immediately beneath it in the Rosseau, Go Home and Britt domains, suggests to Culshaw et al. (1991) that a single unit forms the base of the Parry Sound Domain which lies in thrust contact with underlying units. These retrograded gneisses are not observed in the interior of the Parry Sound Domain.

### Whitestone Suite

Within the Parry Sound shear zone (PSSZ), and associated with the contact of the Lighthouse and McLaren Island gneiss associations, are a series of gabbroic anorthosite bodies (*see* Figures 19.21 and 19.48) including the Parry Island, Whitestone, Arnstein and Raganooter bodies, and their deformed equivalents (Nadeau 1984). Geochemical data on the Whitestone Gabbroic Anorthosite is provided in Mason (1969), Barker (1972), Bright (1990), Marmont and Johnston (1987), Marmont (1988a) and Easton (1990b). Rare earth element geochemistry of the Whitestone Gabbroic Anorthosite is shown in Figure 19.23. Marmont and Johnston (1987) and McRoberts and Tremblay (1991) report geochemical data for the Arnstein and Raganooter anorthosites. van Breemen et al. (1986) report a U-Pb zircon date of 1350±50 Ma from the Whitestone Gabbroic Anorthosite near Parry Sound. Culshaw et al. (1989, 1990, 1991) note that anorthosites along the western margin of the Parry Sound Domain and within the PSSZ can be traced, discontinuously, along that boundary into the shear zone that marks the base of the Moon River Domain.

### STRUCTURE

The PSSZ, as defined by Davidson 1984a, consists of a zone of "lenticular, disrupted, and straight gneiss", roughly coincident with the northern boundary of the Lighthouse gneiss association supracrustal rocks (*see* Figure 19.48), that extends to the belt of mylonites underlying the McLaren Island gneiss association south of Parry Island. Culshaw et al. (1991) note that along Georgian Bay, PSSZ straight gneisses were formed at high grade, exhibiting kyanite in boudin necks in pelitic gneiss. Mylonites characterized by grain-size refinement and porphyroclast shear-sense indicators become increasingly abundant southward across Parry Island, with the latter indicating movement to the north-northwest. These shear-sense indicators appear to have been formed by thrusting during the amphibolite-facies metamorphism.

### RELATIONSHIP BETWEEN THE BRITT AND PARRY SOUND DOMAINS

Much of the Parry Sound shear zone is made of amphibolite-facies straight gneiss. The relationship between these and the Central Britt shear zone and northwest-trending folds, as outlined in Culshaw et al. (1991), is as follows: 1) West of Parry Island, a zone of disharmonic folding in PSSZ straight gneisses passes northwestward directly into CBSZ tectonites. 2) At Killbear Point, southeast-dipping PSSZ tectonites can be projected northwestward over a culmination into a shallowly northwest-plunging fold of the regional northwest-trending synclinorium. Comparably folded foliations elsewhere in the Sand Bay and Ojibway gneiss associations are presumably of the same age. Therefore, the northwest-trending folds in the Britt Domain and the CBSZ cannot predate the basal PSSZ foliations, and foliations coplanar with the latter occur in the southern Britt Domain. This geometry would be expected if the Sand Bay and Objiway gneiss associations were themselves thrust sheets (resulting in the PSSZ-type fabrics), later deformed as the thrust stack evolved (resulting in northwest folds).

### METAMORPHISM

Within the Lighthouse and Armer Bay gneiss associations of the basal Parry Sound Domain there are both kyanite (early) and kyanite-sillimanite-bearing pelites. Peak metamorphic conditions are estimated by Culshaw et al. (1991) at 700° to 750°C and *circa* 8 kb (*see* Table 19.2).

Estimates of peak conditions from the granulites of the Parry Sound Domain are about 800°C and 10 to 11 kb (*see* Table 19.2; Anovitz and Essene 1990). These rocks may have followed a different P-T path from the PSSZ gneisses (Culshaw et al. 1991). Pelites within the granulites contain the stable assemblage, garnet + sillimanite + hypersthene + perthite + quartz + spinel + biotite.

### MINERAL DEPOSITS

A variety of mineral deposits occur in the Parry Sound Domain, in part, reflecting the greater amount of exploration in the area and, in part, the abundance of mafic and supracrustal rocks in the domain. Villard et al. (1984) have summarized data on gold occurrences in the Parry Sound Domain. Gold-copper occurrences are located mainly in the western Parry Sound Domain, in the McLaren Island gneiss association and are associated with the margins of gabbro-diorite bodies in contact with mafic gneisses. Mineralization occurs in concordant to slightly discordant quartz veins generally hosted in the gabbro-diorite rocks.

Several industrial mineral studies have been conducted recently in the Parry Sound Region examining alumina (Marmont and Johnston 1987; Marmont 1988a, 1991), lime (Marmont 1988b, 1991) and flagstone sources (Fouts and Marmont 1989). The best flagstone quarries occur in the PSSZ.

## Moon River Domain

The Moon River Domain (*see* Figure 19.48) has been described as a gently southeast-plunging synformal structure bounded by a continuous inward-dipping tectonite zone (Davidson et al. 1982) that contains long, narrow slivers of highly recrystallized and grain-size reduced meta-anorthosite. Internally, the Moon River Domain contains a variety of pink and light to dark-grey migmatitic gneisses, which commonly contain textures evocative of tectonic origin (Davidson et al. 1982). A long, curved sliver of anorthosite gneiss is present within the core of the synformal structure and helps to define the structure. The anorthosite sliver is highly strained, and to the southeast it becomes a discontinuous zone of anorthositic blocks. Migmatitic gneisses of the Moon River Domain are at amphibolite facies (Davidson 1984b). Dickin and McNutt (1989b) report *circa* 1.4 billion-year-old Nd-model ages from orthogneisses and paragneisses from the Moon River Domain. This is similar to model ages from the Parry Sound Domain, and some 0.4 billion years younger than Nd model ages typical of the Algonquin Terrane (Dickin and McNutt 1989b).

Culshaw et al. (1989, 1990) have mapped the Moon River in some detail (*see* Figure 19.48) and have redefined the northwestern boundary of the domain. The dotted line in Figure 19.48 marks the northwestern boundary of the domain as originally defined by Davidson et al. (1982). Rocks in this area previously included in the Moon River Domain have been assigned to the Parry Sound Domain by Culshaw et al. (1989, 1990).

## GNEISS ASSOCIATIONS

Culshaw et al. (1989, 1990, 1991) have recognized 2 gneiss associations within the Moon River Domain. The structurally lowest unit, the Blackstone Lake gneiss association, consists of variably migmatitic amphibolite-facies granitic to granodioritic orthogneiss with a folded, stratiform, foliation. This orthogneiss, in places, lies with structural discordance on top of variably retrograde granulites of the Parry Sound Domain. No older metamorphism has been detected (Culshaw et al. 1991). Meta-anorthosite, gabbro and ultramafic bodies are present in this association and are similar to the Whitestone Suite within the Parry Sound Domain.

The upper unit, the Moon River gneiss association, has a more varied composition, containing orthogneiss,

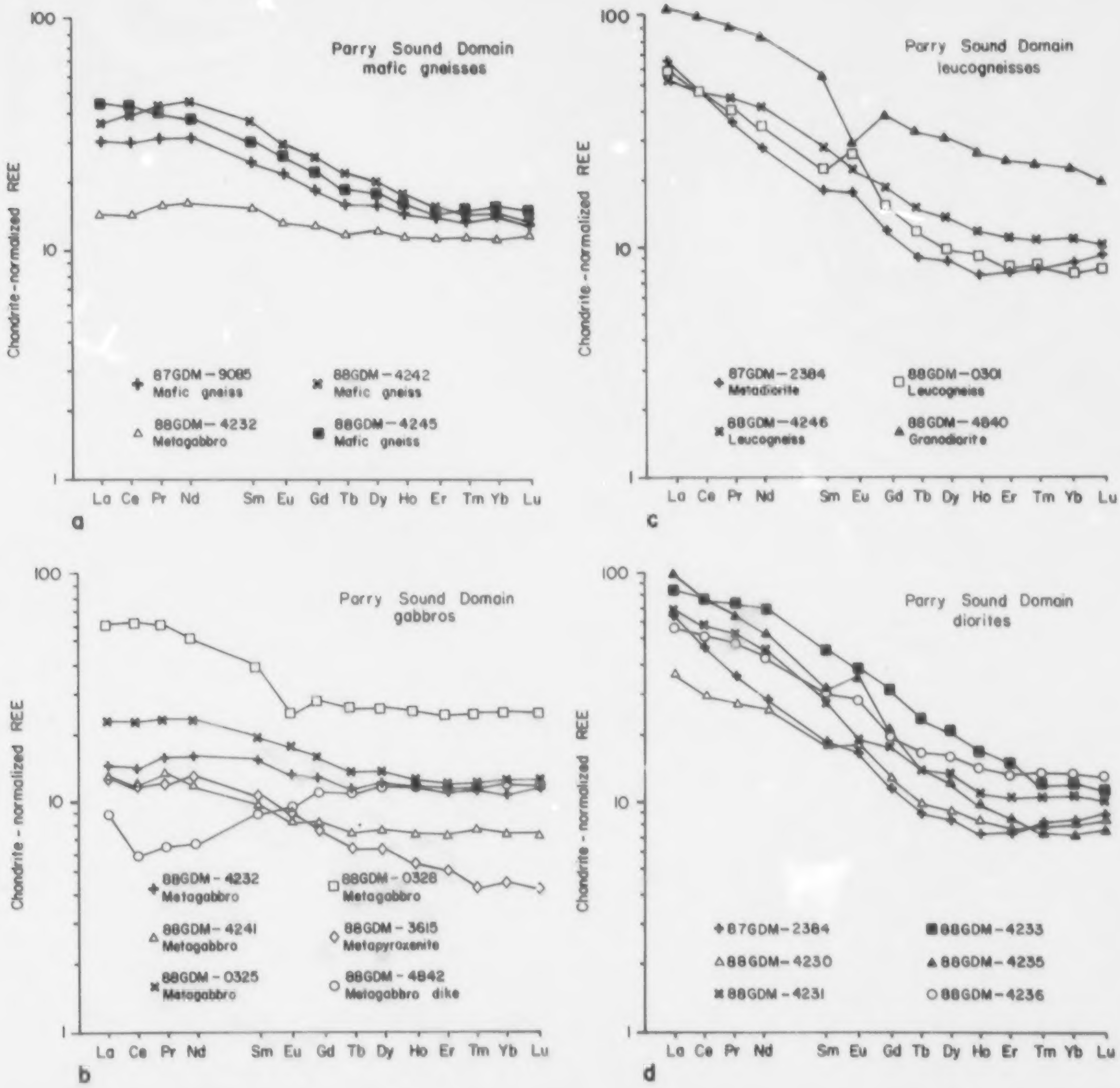


**Figure 19.49.** Rare earth element geochemistry of supracrustal and metaigneous rocks from the Parry Sound Domain. a) Mafic gneisses, possibly derived from mafic volcanic rocks, McLaren Island gneiss association. b) Metagabbros, McLaren Island gneiss association. c) Leucogneiss, Moon Island gneiss association. d) McKellar-type metadiorites, Moon Island gneiss association.

voluminous pink leucogneiss and minor metasupracrustal rocks. Like the Blackstone Lake assemblage, from which it is separated by a thin, attenuated and overturned extension of Parry Sound Domain rocks (*see* Figure 19.48), it is at amphibolite facies. Along its southwestern boundary, the Moon River gneiss association lies discordantly on the Go Home Domain. Culshaw et al. (1991) interpret the boundary of the Moon River and Go Home domains as thrust surfaces on account of discontinuities in lithologic, diking and metamorphic characteristics across the boundary and the fabric of the bounding tectonites (cf., Davidson et al. 1982). Scattered metabasite bodies of coronitic affinity also occur along this boundary.

### MINERAL DEPOSITS

No major mineral deposits have been found to date in the Moon River Domain. Uraniferous pegmatite dikes are abundant near Blackstone Lake, however, ore grades are low.

## Seguin Domain

### GEOLOGY

The Sequin Domain has been described as a gently southeast-plunging synformal structure bounded by a continuous inward-dipping tectonite zone (Davidson et al. 1982, 1985; Culshaw et al. 1983; Macfie and Dixon 1990). It is underlain by layered, medium-grained, migmatitic gneisses of unknown origin that enclose, in the constricted part of the synform, charnockitic, granitic and granodioritic orthogneiss masses. Small blocks and slivers of metamorphosed mafic rock occur throughout this domain (Davidson et al. 1985). Except for the orthogneiss masses, protoliths in this domain are usually indecipherable (Davidson et al. 1985). Orthogneisses from the Seguin Domain have yielded U-Pb zircon ages of 1450 and 1420 Ma (Nadeau and van Breemen 1989) and show no evidence of inheritance. Dickin and McNutt (1989b, 1990) report *circa* 1.4 billion-year-old Nd-model ages from paragneisses from the Seguin Domain. These are similar to model ages from the Parry Sound and Moon River domains, and some 0.4 billion years younger than Nd-Sm model ages typical of the Algonquin Terrane (Dickin and McNutt 1989b). The magnetic background of the Seguin Domain is markedly low relative to the Novar Domain and reflects the geometry of the Sequin synform.

The abundant quartzofeldspathic, migmatitic gneisses of the Sequin Domain are compositionally fairly homogeneous at outcrop and map scales, with 25 to 30% leucosome; neither protolith or paleosome can be discerned (Davidson et al. 1985). The mesosome is commonly medium-grained, equigranular, medium-grey, granodioritic in composition and contains 15 to 25% hornblende and biotite, sometimes with minor red garnet. Small masses of pink, leucocratic biotite granitoid mesosome also occur locally (Davidson et al. 1985). In sections parallel to the lineation, the dominant mesostructures in the leucosome would be described as schlieren and stromatic *after* Mehnert (1968), whereas in sections perpendicular to the lineation, they are phlebitic or folded (Davidson et al. 1985).

### STRUCTURE

Planar structures, namely, mineral foliation and the preferred orientation of leucosome, are well developed in the neck of the domain (*see* Figure 19.48), where they parallel lithologic contacts and along the margins of the domain (Davidson et al. 1985). In the core of the domain, shallow southeast-dipping enveloping surfaces of complex folds, including sheath folds, with fold axes nearly parallel to the mineral lineations are common (Davidson et al. 1985). In the neck, shallow northeast-plunging linear features are locally developed.

Nadeau and van Breemen (1989, 1990) have dated a late synthrusting pegmatite from the base of the Seguin Domain at 1097±3 Ma, suggesting the main period of northwestward-directed thrusting occurred at that time in the Seguin region. However, other late syntectonic pegmatites from the footwall of the Seguin Domain have yielded zircon ages of 1080±1 Ma, 1046±2 Ma and 1027±2 Ma (Nadeau and van Breemen 1990), suggesting renewed deformation at higher structural levels reactivated the lower parts of the thrust stack at these times.

### METAMORPHISM

East of Skeleton Lake, the monotonous migmatites merge into compositionally similar, but less migmatitic gneisses exhibiting dictyonitic mesostructures (Mehnert 1968), and in places, the gneisses grade into pink and grey granitic to granodioritic augen orthogneisses (Davidson et al. 1985). The migmatites and orthogneisses enclose wedge-shaped masses of buff-brown to dark green, medium-grained, hypersthene-bearing charnockitic orthogneiss that contain up to 15% mafic minerals (Davidson et al. 1985). Davidson et al. (1985) suggest that granulite-facies metamorphism and the final foliation in these rocks may predate the development of the Seguin synform and domain emplacement. Nadeau and van Breemen (1990) have dated the emplacement of the Seguin Domain at 1097±3 Ma, thereby implying that granulite-facies metamorphism occurred prior to this time. Nadeau and van Breemen (1990) also report U-Pb monazite blocking ages of *circa* 1065 Ma from the Seguin Domain, suggesting that metamamorphic temperatures greater than *circa* 700°C persisted for at least 35 million years after the tectonothermal event that resulted in the stacking of the Seguin Domain on the Parry Sound Domain.

### MINERAL DEPOSITS

No major mineral deposits have been found to date in the Seguin Domain.

# CENTRAL METASEDIMENTARY BELT BOUNDARY ZONE

The Central Metasedimentary Belt Boundary Zone (CMBBZ) (Davidson et al. 1982, 1984; Hanmer and Ciesielski 1984; Hanmer et al. 1985; Hanmer 1988) is a north- to northeast-trending zone of cataclastic and tectonically disrupted quartzofeldspathic gneisses that separate rocks of the Central Gneiss Belt from the Central Metasedimentary

Belt. Most rocks within the CMBBZ are tectonites, including transposed, straight and porphyroclastic gneiss (*see* Tectonite Nomenclature; *see also* Figures 19.25, 19.26, 19.27, 19.30, 19.31, 19.32 and 19.33). Along most of its length, the tectonites dip shallowly to the south-southeast at 20° to 40°.

Barlow (1899) and Adams and Barlow (1910) recognized that the northern boundary of the Grenville marble belt was a zone of tectonism. Subsequently, later investigators proposed that the base of the Central Metasedimentary Belt was marked by a zone of clastic metasedimentary rocks resting unconformably on rocks of the Central Gneiss Belt (e.g., Appleyard 1974b; Schwerdtner and Lumbers 1980; Lumbers 1982). As noted above, the tectonic interpretation has been reprised.

In the Minden area, the CMBBZ is a 2 to 6 km wide tectonite zone (Figure 19.50). Here, in the westernmost part of the CMBBZ, the tectonites can be shown to be derived from Fishog Domain gneisses (Easton 1987d, 1990a). In the core of the CMBBZ, protolith is indeterminate. The eastern contact with the CMBBZ is sharp, with granitoid straight gneiss in fault contact with marble tectonic breccia of the Denna Lake Structural Complex (Easton 1990a).

North of Haliburton, the CMBBZ changes orientation from north to northeast, and several large thrust complexes of *circa* 1340 million-year-old tonalite gneisses are interleaved with the tectonites (Hanmer 1988, 1989). McEachern et al. (1990) also report several large thrust sheets in the CMBBZ in the Pembroke area. Between Haliburton and Minden, the boundary with the Bancroft Terrane is difficult to define because of interleaved marble tectonic breccia and thrust nappes.

McEachern (1990) considered large areas of the Bancroft Terrane to represent a Central Metasedimentary Belt Boundary Tectonic Zone, with the most northerly zone of tectonites (coarse clastic sequence *of* Lumbers 1982) representing a boundary thrust. The usage proposed herein is to define the western and northernmost zone of tectonites (generally less than 5 km wide) as the CMBBZ. The marble tectonic breccias and the interleaved thrust sheets are considered part of the Bancroft Terrane (i.e., the Central Metasedimentary Belt). This usage is in accordance with historical designation and approximates the boundary between the Central Gneiss and Central Metasedimentary belts as defined by Wynne-Edwards (1972), Ayres et al. (1971b) and Lumbers (1982). This usage does not preclude the Bancroft Terrane being a tectonic zone, and in fact, preliminary results from the Abitibi–Grenville LITHOPROBE seismic reflection line suggest that this is indeed the case, as the CMBBZ and the Bancroft Terrane are seismically continuous (White et al. 1991a).

Zircon ages from syntectonic pegmatites from the CMBBZ in the Haliburton area constrain movement of thrust sheets in the CMBBZ to an episode of about 1065±15 to 1060±6 Ma (van Breemen and Hanmer 1986). McEachern (1990) dated strongly zoned zircons from syntectonic pegmatites in the eastern CMBBZ and obtained ages of $1181^{+60}_{-18}$ and $1189^{+35}_{-11}$ Ma from Killaloe and 1053±3 and $1056^{+20}_{-3}$ from the Pembroke and Papineau Lake areas, respectively. The older ages reflect the initiation of thrusting, with subsequent reactivation of the shear zone at 1065 to 1060 Ma. In addition, a syntectonic pegmatite from Killaloe was dated at 1029±14 Ma, and a posttectonic pegmatite yielded an age of 1010 Ma (van Breemen and Hanmer 1986). These ages, in conjunction with titanite ages of *circa* 1040 to 1030 Ma from the Bancroft Terrane (Mezger et al. 1991), suggest thrusting was complete prior to 1030 Ma. In fact, Mezger et al. (1991) suggest that late movement along the ductile shear zones in the CMBBZ reflects postorogenic collapse of the thrust stack, consequently, construction of the thrust stack was probably complete by *circa* 1045 Ma. This 1045 to 1030 million-year-old activity overlaps a period of alkalic metasomatism along the CMBBZ and within the Bancroft Terrane, and the relationship between the 2 events is unclear. The ages of movement from the CMBBZ are, however, generally consistent with ages from the Huntsville area (Nadeau and van Breeman 1990), which indicate early (1180 to 1140 Ma) and late (1070 to 1060 Ma) movement on ductile Central Gneiss Belt thrusts (*see* Figure 19.19).

## CENTRAL METASEDIMENTARY BELT

### Tectonic Divisions

The Central Metasedimentary Belt in Ontario can be divided into 5 lithotectonic terranes, as illustrated in Figure 19.51. The Elzevir Terrane can be further divided into 3 domains, the Harvey-Cardiff Arch, and the Belmont and Grimsthorpe domains. These domains may represent additional terranes, however, they are sufficiently similar that they are considered here as part of the Elzevir Terrane. The terrane division is similar to that proposed by Moore (1982) (*see* Figure 19.7e) except for the addition of the Mazinaw Terrane and revisions to the Bancroft–Elzevir terrane boundary. Figure 19.52 and Table 19.5 highlight some of the key differences between the various terranes. Figure 19.51b illustrates the possible stacking order and relationships between the different terranes.

### Organization of this Section

Previous overviews of Central Metasedimentary Belt geology (cf., Lumbers 1967a, 1967b; Sangster and Bourne 1982; Carter 1984; Easton et al. 1986) emphasized common features across the Central Metasedimentary Belt. This review emphasizes the differences between parts of the Central Metasedimentary Belt by illustrating the differences between apparently similar regions, for example stratigraphy or mineral deposits. This approach allows the development of more site specific mineral deposit models and a better understanding of the history and tectonic development of the Central Metasedimentary Belt. This reductionist approach does, however, present a much more complicated picture of the Central Metasedimentary Belt, which may be confusing to the Grenville neophyte.

Consequently, this part of the chapter is divided into 3 sections: !) an overview of the broader aspects of Central

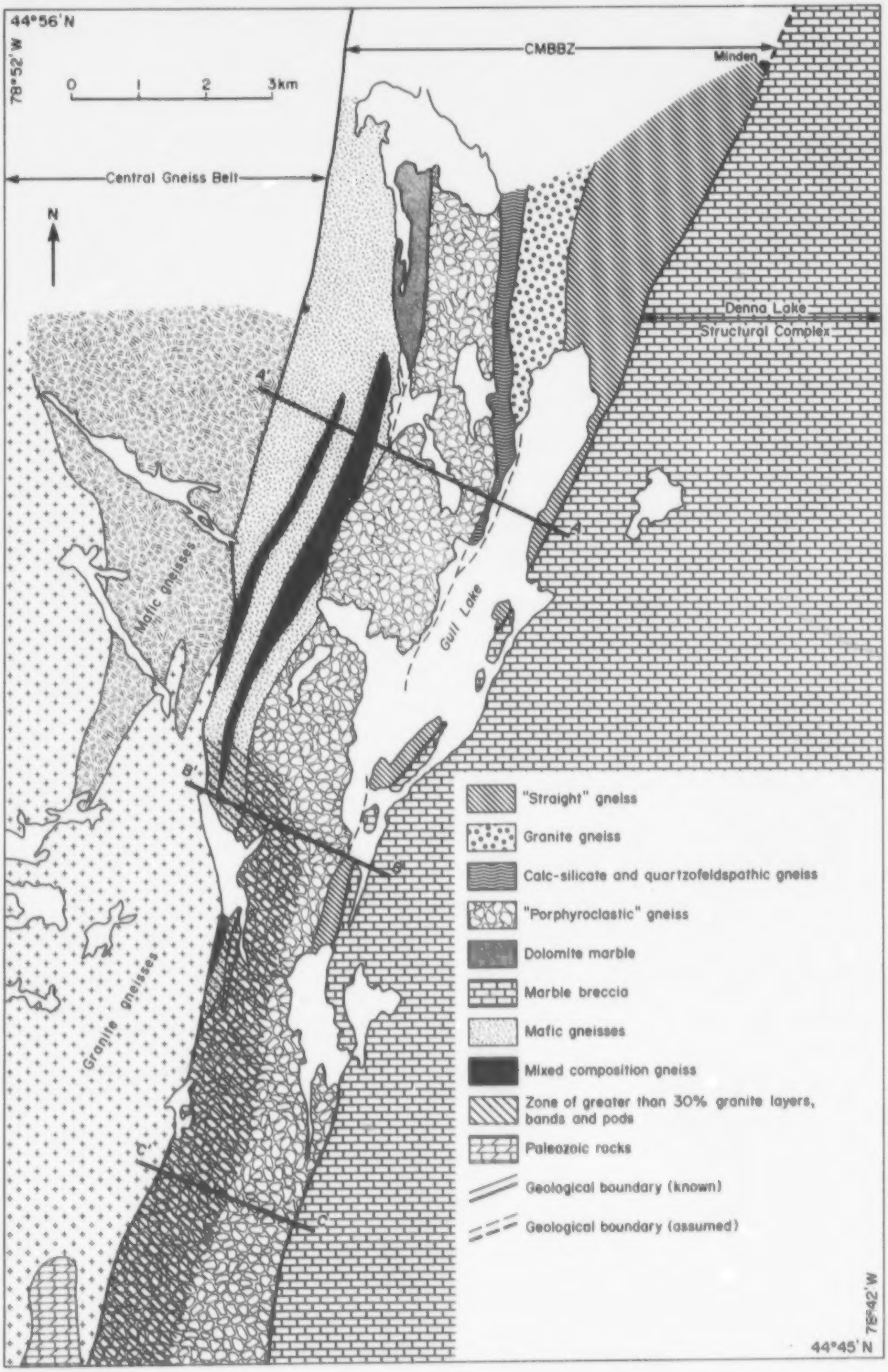

44°56'N
78°52'W
0 1 2 3km
CMBBZ
Minden
Central Gneiss Belt
N
Denna Lake
Structural Complex
A
A'
B
B'
C
C'
Mafic gneisses
Gull Lake
Granite gneisses
"Straight" gneiss
Granite gneiss
Calc-silicate and quartzofeldspathic gneiss
"Porphyroclastic" gneiss
Dolomite marble
Marble breccia
Mafic gneisses
Mixed composition gneiss
Zone of greater than 30% granite layers, bands and pods
Paleozoic rocks
Geological boundary (known)
Geological boundary (assumed)
78°42'W
44°45'N

**Figure 19.58.** a) Lithotectonic subdivisions of the Central Metasedimentary Belt Boundary Zone (CMBBZ) in the Minden area. b) Rock types present in 3 sections across the CMBBZ in the Minden area. For location of sections *see* (a) (Easton 1986b, 1990a).

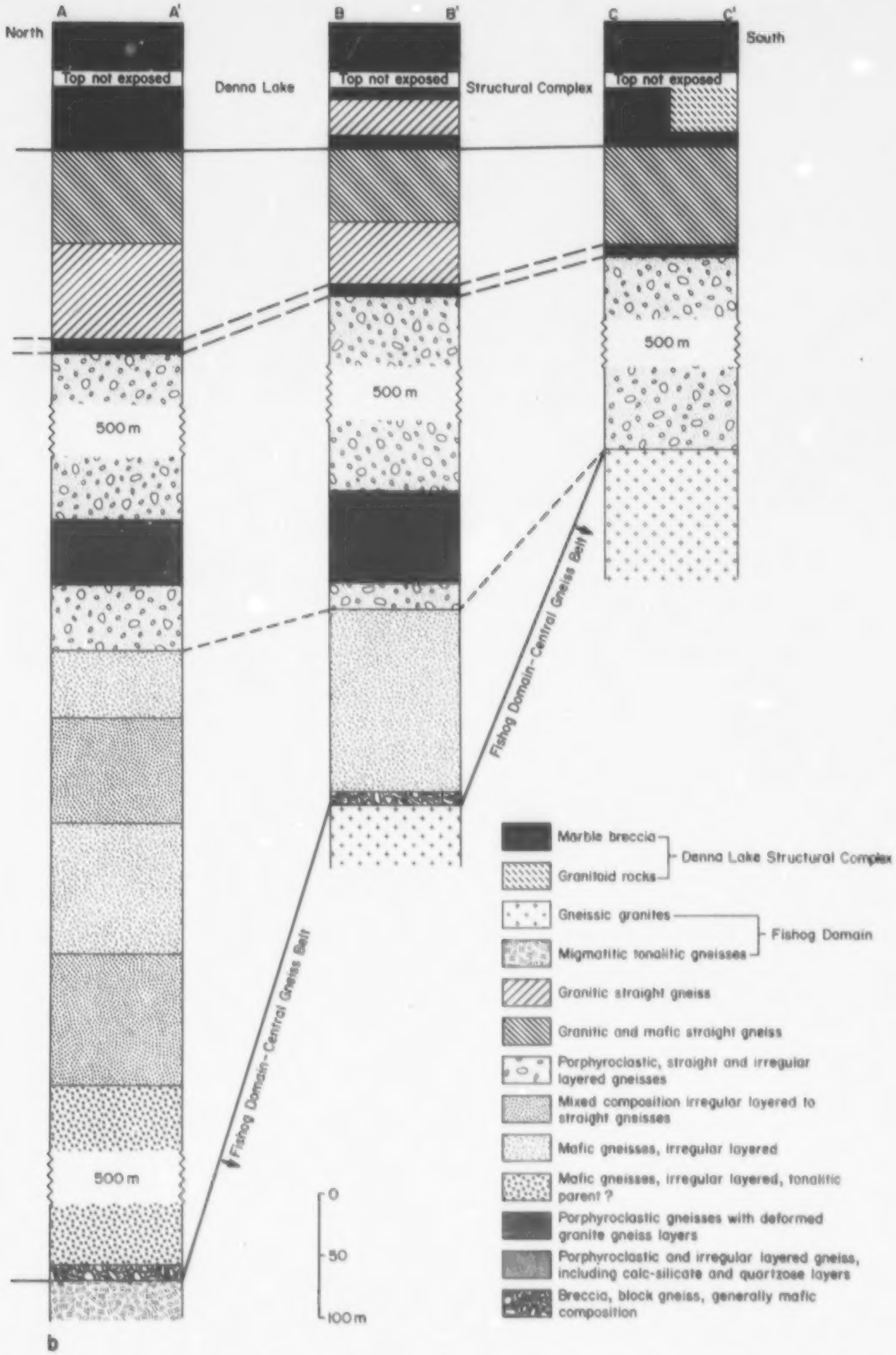

A
A'
B
B'
C
C'
North
South
Top not exposed
Top not exposed
Top not exposed
Denna Lake
Structural Complex
500 m
500 m
500 m
500 m
Fishog Domain–Central Gneiss Belt
Fishog Domain–Central Gneiss Belt
Marble breccia
Granitoid rocks
Denna Lake Structural Complex
Gneissic granites
Migmatitic tonalitic gneisses
Fishog Domain
Granitic straight gneiss
Granitic and mafic straight gneiss
Porphyroclastic, straight and irregular layered gneisses
Mixed composition irregular layered to straight gneisses
Mafic gneisses, irregular layered
Mafic gneisses, irregular layered, tonalitic parent ?
Porphyroclastic gneisses with deformed granite gneiss layers
Porphyroclastic and irregular layered gneiss, including calc-silicate and quartzose layers
Breccia, block gneiss, generally mafic composition
0
50
100 m
b

**Figure 19.50.** a) Lithotectonic subdivisions of the Central Metasedimentary Belt Boundary Zone (CMBBZ) in the Minden area. b) Rock types present in 3 sections across the CMBBZ in the Minden area. For location of sections *see* (a) (Easton 1986b, 1990a) .

Metasedimentary Belt geology, such as styles of volcanism, carbonate sedimentation, stratigraphic nomenclature, plutonic styles and metallogeny; 2) the geology of individual domains including their mineral deposits and contrasts and similarities with adjacent areas; and 3) a tectonic synthesis of the Central Metasedimentary Belt included in the tectonic summary section. Information on mineral deposits is found in all 3 sections and is summarized by Easton and Fyon (this volume).

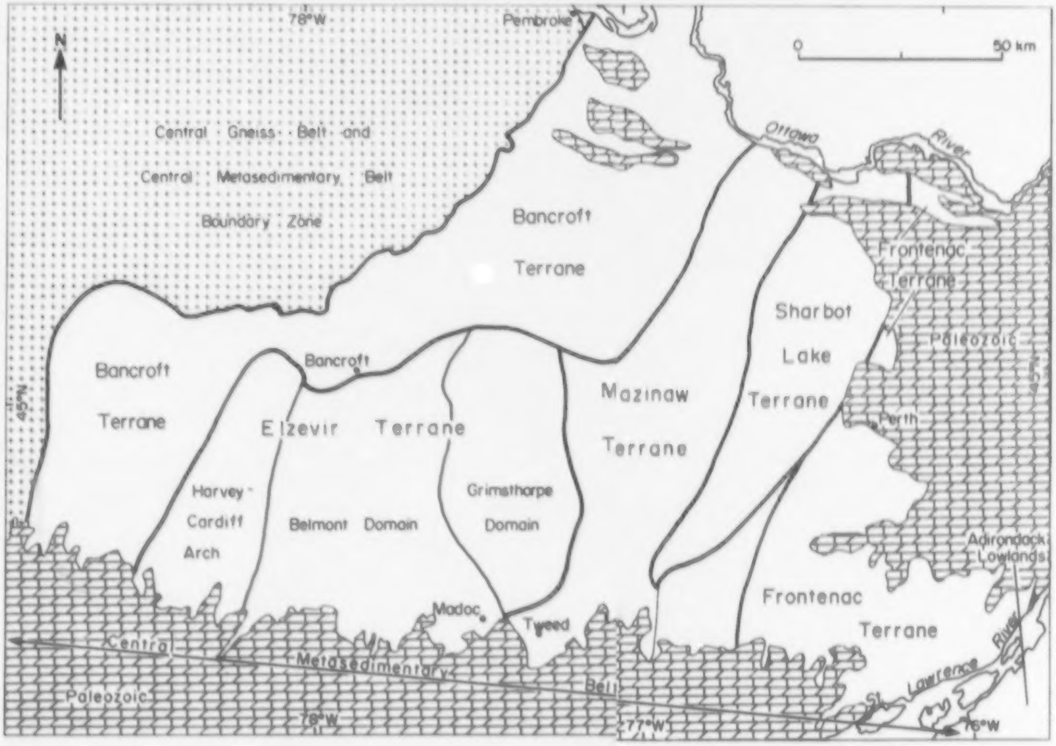
## Stratigraphic Overview

As the Grenville Province was among the first part of the Canadian Shield mapped in Canada (Logan 1863, 1866), a considerable body of literature exists with regards to the stratigraphy of the Grenville Province (Table 19.6). The stratigraphic nomenclature used in this chapter is considerably revised from existing stratigraphic schemes and is summarized in Figure 19.53 and in Table 19.7 with more detail available in Tables 19.8 to 19.14. In order to properly understand the literature on the Grenville Province, it is necessary to provide a historic component to this section beyond that given in Table 19.6.

### GRENVILLE SUPERGROUP

The Grenville Supergroup (formerly series, Group) is a sequence of stratified metavolcanic and metasedimentary rocks exposed in the Central Metasedimentary Belt of eastern Ontario and western Quebec (Lumbers 1967a, 1967b, 1969). The term has been applied also to any supracrustal rocks of supposed Mesoproterozoic age within the Grenville Province (e.g., Parry Sound Domain supracrustals). The term dates back to Logan (1863), who coined the term "Grenville Series" for an assemblage of quartzite, marble and sillimanite-garnet gneiss exposed near the village of Grenville, Quebec. Subsequently, its use has been extended over large areas of the Grenville Province, and the term adapted to modern stratigraphic nomenclature (e.g., NACSN 1983). Contributors to the Royal Society of Canada volume on "The Grenville Problem" (Thomson 1956) devoted extensive discussion to these stratiform rocks; the need was expressed to avoid time-stratigraphic connotations and overdependence on correlations based on Logan's "trinity" and instead to identify distinctive lithologic assemblages. Today, the term is still ill-defined, in part, because of overdependence on regional correlations not substantiated by detailed stratigraphic analysis.

Hewitt and James (1956) suggested subdivisions later formalized by Lumbers (1967a, 1969; *see* Table 19.11) in the Bancroft–Madoc area, primarily in Tudor and Cashel townships. Wynne-Edwards (1972) and Moore and Thompson (1972) elevated these supracrustal rocks from "Group" to "Supergroup" (*see* Table 19.6). Wynne-Edwards (1972) considered the succession to be older than *circa* 1.65 Ga on


78°W
Pembroke
N
0
50 km
Central Gneiss Belt and
Central Metasedimentary Belt
Boundary Zone
Ottawa
River
Bancroft
Terrane
Frontenac
Terrane
Paleozoic
Sharbot
Lake
Terrane
Bancroft
Bancroft
Terrane
45°N
Mazinaw
Terrane
Elzevir Terrane
Perth
Harvey-
Cardiff
Arch
Belmont Domain
Grimsthorpe
Domain
Adirondack
Lowlands
Madoc
Tweed
Frontenac
Terrane
Central
Metasedimentary
Belt
Paleozoic
St. Lawrence
River
78°W
77°W
76°W

a

**Figure 19.51a.** Lithotectonic terranes and domains of the Central Metasedimentary Belt in Ontario as used in this report.

the grounds that xenoliths of quartzite and marble are found in anorthosites of that age. Moore and Thompson (1980) pointed out that the Grenville Supergroup (including the type locality) is no older than 1.3 Ga. Moore and Thompson (1980) also removed the Flinton Group as it unconformably overlies the Grenville Supergroup, although according to the North American Stratigraphic Code (NACSN 1983), supergroups can include major unconformities. Recent U-Pb dating of detrital zircons from the Flinton Group (Kinsman and Parrish 1990) support Moore and Thompson's (1980) suggestion and has laid the debate on the age of the Flinton Group, at least in its type area, to rest.

Moore (1986) restricted the term Grenville Supergroup to the contiguous succession in the Central Metasedimentary Belt which was deposited in the interval 1.3 to 1.2 Ga. Wynne-Edwards (1967b, 1972) recognized that the Grenville Supergroup stratigraphy within the Frontenac Axis (Terrane) differs from that of the Bancroft Terrane, presaging subsequent revisions to stratigraphic nomenclature described below.

## FLINTON GROUP ("HASTINGS SERIES")

In 1866, Logan introduced the term "Hastings Series" to describe rocks which he considered to be less altered equivalents of the "Grenville Series." Vennor (1870) proposed that the "Hastings Series", mainly metaconglomerates and meta-arenites, was younger than the Grenville "series." As summarized in Table 19.6, this led to an ongoing debate over the stratigraphic relationships between the Grenville and Hastings "series." This debate did not cease even after Moore and Thomson (1972, 1980) and Thompson (1972) established that the "Hastings series", which they renamed the Flinton Group, locally unconformably overlies the Grenville Supergroup and lacked a deformational event present in the Grenville Supergroup. Moore and Thompson (1972, 1980) also included within the Flinton Group a sequence of black sulphidic shales, calcitic and dolomitic marbles and marble breccias and conglomerates (Myer Cave Formation) and calcareous and clastic rocks (Fernleigh Formation), units which had not previously been considered part of the "Hastings Series."

There is now consensus that the Flinton Group unconformably overlies older supracrustal and metaplutonic rocks (Smith 1958; Moore and Thompson 1972, 1980; Thompson 1972; Easton and Ford 1991). Recent U-Pb dating of detrital zircons of the Flinton Group (Kinsman and Parrish 1990) support Moore and Thompson's (1980) separation of the Flinton Group from the Grenville Supergroup and the fact that it is significantly younger (by almost 100 million years) than the Grenville Supergroup.

## Revised Stratigraphic Framework

The stratigraphic framework of the Central Metasedimentary Belt was established before mapping, structural,

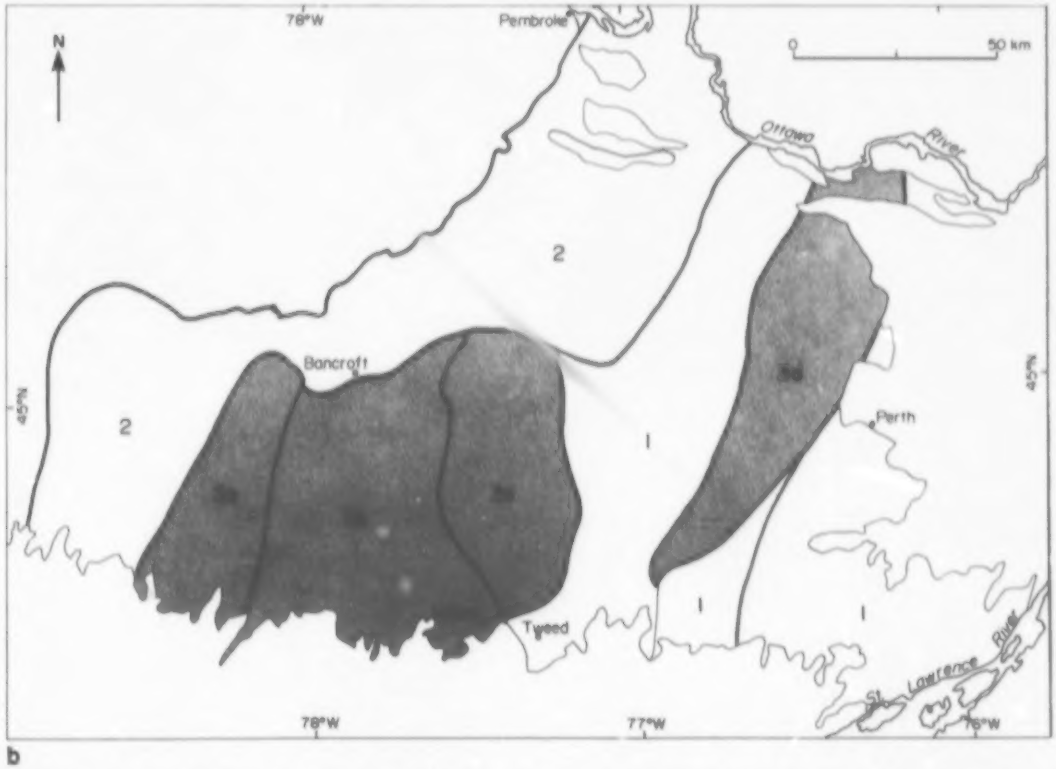


**Figure 19.51b.** Sketch showing possible genetic links and stacking or accretion relationships between terranes and domains in the Central Metasedimentary Belt.

geophysical and chronometric studies revealed its tectonic and chronometric diversity. This prior stratigraphic framework did not recognize the provinciality now postulated with respect to the distribution of various stratigraphic units (e.g., Davidson 1986a). Consequently, it is necessary to revise the stratigraphic nomenclature of the Central Metasedimentary Belt to reflect current knowledge.

To characterize the lithologic diversity across the Central Metasedimentary Belt, a series of stratigraphic sections are shown in Figure 19.54. This figure illustrates the difficulties in any attempt to postulate regional correlations in the Central Metasedimentary Belt.

The revised stratigraphy is illustrated in Figures 19.53a and 19.54 and in Table 19.7; stratigraphic units are briefly described below. Only a minimum of new units have been created. Outdated and abandoned stratigraphic terms are listed in Table 19.8. The purpose of this revision is to clarify stratigraphic correlations within the Central Metasedimentary Belt.

## FRONTENAC SUPERGROUP (NEW TERM)

The supracrustal rocks of the Frontenac Terrane (Axis) are herein removed from the Grenville Supergroup and placed in the Frontenac Supergroup (*see* Figures 19.52 and 19.54 and Table 19.7). These rocks differ from the Grenville Supergroup in their lack of volcanic and volcaniclastic rocks and the abundance of quartzite units. The stratigraphy of this area has not been revised since the work of Wynne-Edwards (1967a, 1967b), therefore, no formal groups or formations are proposed here for the Frontenac Supergroup, several could be established (*see* Wynne-Edwards 1967a, 1972; Figure 19.55). The type area for the Frontenac Supergroup was defined by Wynne-Edwards (1967a) as the Westport area. The abundance of quartzites and siliceous clastic metasedimentary rocks increases to the east and southeast in the Frontenac Terrane (Wynne-Edwards 1967a; Lumbers et al. 1990).

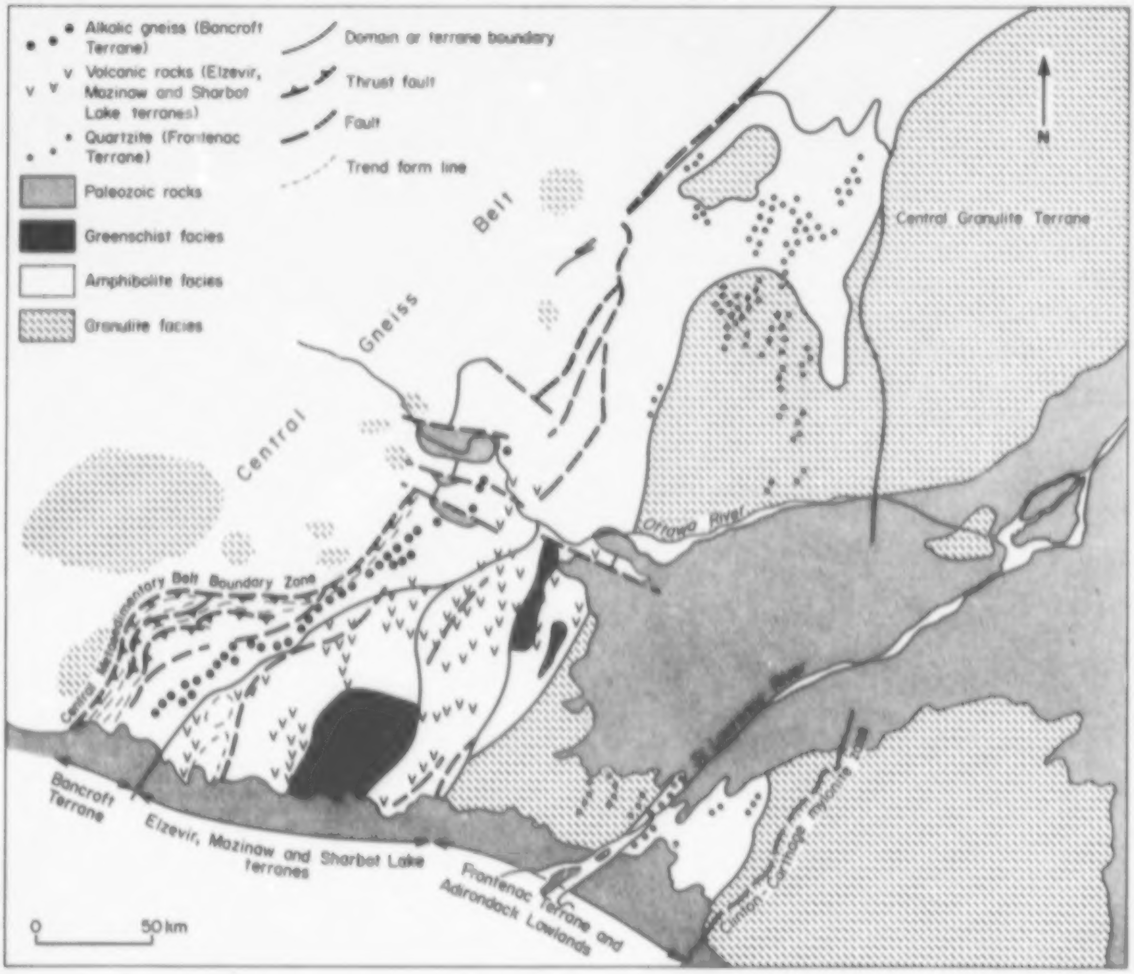

Alkalic gneiss (Bancroft Terrane)
Volcanic rocks (Elzevir, Mazinaw and Sharbot Lake terranes)
Quartzite (Frontenac Terrane)
Paleozoic rocks
Greenschist facies
Amphibolite facies
Granulite facies
Domain or terrane boundary
Thrust fault
Fault
Trend form line
N
Central Gneiss Belt
Central Granulite Terrane
Ottawa River
St Lawrence River
Central Metasedimentary Belt Boundary Zone
Bancroft Terrane
Elzevir, Mazinaw and Sharbot Lake terranes
Frontenac Terrane and Adirondack Lowlands
Colton - Carthage mylonite
0
50 km

**Figure 19.52.** Simplified map showing lithologic differences between the terranes within the Central Metasedimentary belt. *Modified from* Davidson 1986a.

Correlation between the Grenville Supergroup in the Bancroft and Elzevir terranes and the Frontenac Supergroup cannot be established at this time. The 2 supergroups may be roughly time-correlative, even if they were deposited in different geographic settings. As outlined below (*see* Tectonic Summary), the Elzevir Terrane and the Frontenac Terrane may not have been contiguous until *circa* 1160 Ma, and therefore, the sequences may not be correlative. The marbles in both supergroups were tentatively correlated, however, if these rocks have not been transported far (i.e., less than several hundred kilometres) from their site of deposition by later tectonic activity, then paleolatitudes conducive to carbonate deposition existed in the region between 1450 to 1250 Ma (Irving 1979). Thus, a significant (more than 50 million years) age difference may exist between carbonate deposition in the 2 supergroups.

The maximum depositional age for the youngest members of the Frontenac Supergroup, based on detrital zircons from the Upper Marble and Quartzite unit is 1301±13 Ma (Kinsman and Parrish 1990). The youngest detrital zircons from quartzites lower in the section (below the Major Marble unit, *see* Figure 19.55) are 1415±2 Ma in age (Kinsman 1990). The younger age is at least 40 to 50 million years older than the main carbonate sequence of the Grenville Supergroup (*see* Grenville Supergroup). Additional detrital zircons range from 1.5 to 2.6 Ma, with a cluster of ages between 1745 and 1892 Ma. Other evidence for the age of the Frontenac Supergroup comes from the Wellesley Island leucogranite gneiss in New York, which contains metasedimentary xenoliths dated at 1415±6 Ma (McLelland, Chiarenzelli et al. 1988). Marcantonio et al. (1990) report Nd model ages from metasedimentary gneisses of the Frontenac Terrane between 2045 and 1560 Ma, suggesting that paragneisses of the lower part of the Frontenac Supergroup may have been deposited between 1.8 and 1.4 Ga (age of Wellesley Island leucogranite gneiss).

## GRENVILLE SUPERGROUP

The Grenville Supergroup is being retained in its historic context as an all encompassing unit of *circa* 1300 to 1200 million-year-old supracrustal rocks from the bulk of the Central Metasedimentary Belt. However, the Hermon and Mayo groups are considered restricted to the Belmont and, perhaps, part of the Grimsthorpe Domain. Thus, several new domain-specific groups are being established, herein, as illustrated in Figures 19.53a and 19.54 and Tables 19.7 to 19.14. In this section, the stratigraphic framework is defined in broad terms. Detailed descriptions of newly proposed groups are found below in the detailed Central Metasedimentary Belt terrane descriptions.

As outlined in latter sections of this chapter, despite the distinct lithologic, structural and tectonic histories of many of the domains and terranes of the Central Metasedimentary Belt, similarities in the above aspects of the Bancroft, Elzevir, Mazinaw and Sharbot Lake terranes suggest they are genetically related, even if this relationship is poorly understood at the moment. Therefore, a broadly defined Grenville Supergroup is retained.

## BANCROFT TERRANE

### Haliburton Group (New Term)

The term Haliburton Group is proposed for the nontectonically disrupted strata located mainly in the southeastern parts of the Bancroft Terrane. The Haliburton Group consists of calcitic and dolomitic marbles, impure marbles, quartz arenites and minor mafic to intermediate volcaniclastic rocks and para-amphibolite (*see* Table 19.9). Rock types in the Haliburton Group are described in greater detail below in the terrane description. The Haliburton Group includes the Salerno Lake and Tory Hill formations (Bright 1976, 1977; *see* Table 19.9). No additional formations within the Haliburton Group are proposed. Reference sections for the Haliburton Group are provided in Easton (1987a) and Bright (1976, 1977).

Correlation of the Haliburton Group with groups in the Elzevir Terrane is problematic, in large part due to uncertainties in depositional age of the Haliburton Group. Lumbers et al. (1990) proposed a new supracrustal sequence in the Minden area, the Haliburton sequence (roughly equivalent to the Haliburton Group), based on reported intrusive relations with tonalite gneisses in the region, such as the $1344^{+93}_{-32}$ million-year-old Dysart Tonalite Gneiss (van Breemen and Hanmer 1986). Hanmer (1988, 1989) and Easton (1990a, 1990c), among others, have suggested that the Dysart Suite was structurally emplaced, and that therefore, the Haliburton Group may have been deposited at roughly the same time as the Mayo Group in the Elzevir Terrane. Regardless, the dominance of shallow-water carbonate and siliciclastic facies in the Haliburton Group *versus* the deeper water carbonate and siliciclastic facies in the Mayo Group validates distinction of the Haliburton Group.

The Bancroft Terrane also includes the Denna Lake Structural Complex (*see* Tables 19.7 and 19.9; Easton 1983, 1990a), which consists of tectonically disrupted Haliburton Group strata (Easton 1983, 1985b, 1987a, 1987e, 1990a). The Denna Lake Structural Complex is younger than the Haliburton Group. It is not universally possible to distinguish areas underlain by Haliburton Group strata and structural complexes equivalent to the Denna Lake Structural Complex. Perhaps the central and northern Bancroft Terrane is best described as several structural complexes rather than as groups. In the interim, however, the term Haliburton Group will serve as a useful reconnaissance mapping tool until the stratigraphy of the Bancroft Terrane is better established.

## ELZEVIR TERRANE

### Harvey–Cardiff Arch

Bright (1976, 1977, 1987a, 1987b) introduced several new formations within the Harvey–Cardiff Arch (*see* Table 19.10) which he considered to be part of the Hermon Group. These include the dominantly metasedimentary Monmouth, Catchacoma and Glamicarde formations and the dominantly metavolcanic Cavendish and Eels Lake formations. These formations are retained, with

**Table 19.5.** Summary chart outlining characteristic features of the Central Metasedimentary Belt terranes.

| Name | Major Supracrustal Rock Types | Volcanic Geochemical Affinities | Age of Volcanism and/or Sedimentation |
|---|---|---|---|
| **Bancroft Terrane** | calcite and dolomite marbles, quartz arenite, quartzofeldspathic gneiss, minor para-amphibolite.<br>Haliburton Group and Denna Lake Structural Complex | volcaniclastic rocks only, alkali basalt and calc-alkalic | >1400 Ma<br><1250 Ma |
| **Elzevir Terrane** | | | |
| Harvey–Cardiff Arch | mafic and felsic metavolcanic rocks, metawackes, metapelite, turbidites, calcite and dolomite marbles<br>Grenville Supergroup | bimodal tholeiitic basalt-dacite-rhyolite | >1300 Ma<br><1250 Ma |
| Belmont Domain | mafic, intermediate and felsic metavolcanic rocks and volcaniclastic sedimentary rocks, siliceous clastic metasedimentary rocks, deep and shallow water calcite and dolomite marble containing stromatolites<br>Grenville Supergroup, Mayo and Hermon groups | 80% bimodal tholeiitic basalt-dacite-rhyolite, minor calc-alkalic andesite and calc-alkalic andesite-rhyolite | >1300 Ma<br>>1260 Ma |
| Grimsthorpe Domain | mafic metavolcanic rocks and volcaniclastic metasedimentary rocks<br>Grenville Supergroup, Canniff and Grimsthorpe groups | tholeiitic | >1350 Ma<br><1250 Ma |
| **Mazinaw Terrane** | mafic, intermediate and felsic metavolcanic rocks, volcaniclastic metasedimentary rocks, siliceous clastic metasedimentary rocks, deep and shallow water calcite and dolomite marbles containing stromatolites, unconformably overlain by conglomerate, arenite, pelite, schist and marbles of 1157 million years in age<br>Mazinaw and Flinton groups | calc-alkalic andesite-rhyolite, minor tholeiitic basalt | >1300 Ma<br><1250 Ma<br>unconformity<br>1150 Ma sedimentation |
| **Sharbot Lake Terrane** | mafic volcanic rocks, minor volcaniclastic and siliceous clastic metasedimentary rocks, dominantly calcite and dolomite marbles, stromatolitic<br>Sharbot Lake Group | tholeiitic | >1300 Ma<br><1250 Ma |
| **Frontenac Terrane** | paragneiss, quartzofeldspathic gneiss, calcite and dolomite marbles, quartz arenite<br>Frontenac Supergroup | no volcanic rocks present | >1800 Ma<br><1180 Ma |

correlations with the revised Hermon Group as illustrated in Figure 19.54.

## Belmont Domain

### HERMON AND MAYO GROUPS (SLIGHTLY REVISED TERMS)

The original stratigraphic sequence for this domain (Lumbers 1967a, 1967b; *see* Table 19.11) is retained, with modifications (*see* Figures 19.54 and 19.56; *see* Table 19.12). The Mayo Group is, in part, a lateral facies equivalent of the Hermon Group (*see* Figure 19.56b) and has been expanded to include most metasedimentary units within the sequence, leaving the Hermon Group as dominantly metavolcanic in character (*see* Table 19.12). Additional formations defined by Bartlett (1983) and Bartlett and Moore (1985) have also been included (*see* Table 19.12). The relationship of the Stoney Lake Metavolcanics of Bright (1988) to existing stratigraphic units (Oak Lake and Burnt Lake formations) is not known, consequently, the Stoney Lake Metavolcanics are not defined as a stratigraphic unit in Table 19.12.

Recent geochronology (Davis and Bartlett 1988; Lumbers et al. 1990) indicates that 2 ages of metavolcanic rocks are present in the domain: an older, predominantly mafic sequence (Tudor Formation *of* Lumbers 1967a, 1967b) deposited between 1290 and 1280 Ma and a younger, compositionally diverse metavolcanic sequence (Burnt Lake, Oak Lake and Turriff formations) deposited between

**Table 19.5.** Summary chart outlining characteristic features of the Central Metasedimentary Belt terranes.

| Major Plutonic Suites Present | Metamorphic Conditions | Major Mineral Deposits | Potential Basement |
|---|---|---|---|
| Dysart, nepheline syenite, anorthosite, Killer Creek, Elzevir, Lavant, Methuen, Skootamatta, fenite-carbonatite-granite pegmatite | middle to upper amphibolite facies,<br>2 regional metamorphic episodes | stratiform Zn, metavolcanic Fe, Mo and U deposits, flagstone, dolomite, graphite, corundum, feldspar, mica, mineral collecting | tonalitic, ca.1350 Ma or older |
| Elzevir, Methuen, fenite-carbonatite-granite pegmatite | middle to upper amphibolite facies | Mo and U deposits | unknown |
| nepheline syenite, Elzevir, Lavant, Methuen, Skootamatta | upper greenschist to lower amphibolite facies | sulphide deposits, iron formation, contact metasomatic Fe deposits, Au deposits, talc, wollastonite, calcite and dolomitic marble | unknown |
| Killer Creek, Elzevir, Skootamatta | upper greenschist to lower amphibolite facies | sulphide deposits, Cu-PGE-Cr magnetite deposits, talc, building stone | gabbroic <1290 million years in age, represented by Canniff Complex |
| anorthosite, Killer Creek, Elzevir, Methuen, granite pegmatite | lower to upper amphibolite facies<br>2 regional metamorphic events, second event isotropic with large porphyroblasts | Au deposits, sulphide deposits, sillimanite, garnet, mica, kyanite, stratiform Zn | Grimsthorpe domain-type basement to supracrustal rocks |
| Lavant, Methuen, Skootamatta? | lower greenschist to middle amphibolite facies | Au deposits, calcite marble, stratiform Zn | unknown |
| anorthosite, Gananoque, Skootamatta, granite pegmatite | granulite facies | no significant deposits | >1400 Ma, possibly as old as 1800 Ma, granitic to granodioritic |

1260 and 1248 Ma (*see* Table 19.11). Consequently, the Hermon Group has been revised in Table 19.12 to reflect these 2 sequences. An unnamed Basal Group is created containing the older Tudor and Cordova Lake metavolcanic rocks. The Hermon Group is redefined to include only metavolcanic rocks of the younger sequence.

## Grimsthorpe Domain

### CANNIFF COMPLEX AND GRIMSTHORPE GROUP (NEW TERMS)

Two new stratigraphic terms are proposed. The volumetrically minor, poorly exposed Canniff Complex includes deformed mafic metavolcanic rocks and protomylonitic gabbros that may be among the oldest rocks in the Central Metasedimentary Belt (Easton and Ford 1990a, 1990b, in prep.). Type areas are indicated in Table 19.13.

Also present within the Grimsthorpe Domain is the Kaladar Complex (Chappell 1978; Wolff 1982a), which includes rocks of the Canniff Complex and the Grimsthorpe Group, as well as intrusive rocks of the Killer Creek Suite. Because of this extreme lithologic diversity and the wide variation in age of the constituent units, it is proposed that the term Kaladar Complex be abandoned as a formal stratigraphic term.

The bulk of the supracrustal rocks in the Grimsthorpe Domain consist of dominantly mafic volcaniclastic metasedimentary rocks, rusty-weathering schists, and minor pelite

**Table 19.6.** History of stratigraphic nomenclature in the Central Metasedimentary Belt in Ontario.

| **Reference** | **Stratigraphic name and interpretation** |
|---|---|
| Logan (1846) | Grenville Series north of the Ottawa River. |
| Logan (1866) | Hastings Series is a less altered equivalent of the Grenville Series. |
| Vennor (1870) | Hastings Series is younger than the Grenville Series. |
| Adams et al. (1907) | Hastings Series is part of the Grenville Series, but the name should be retained to indicate the difference in metamorphic grade; the name Hastings Series is applied to all relatively low-grade supracrustal rocks in Ontario. |
| Miller and Knight (1914) | Hastings Series is unconformable on the Grenville Series in southeast Ontario; Grenville and Hastings series are correlated with the Keewatin and Temiskaming series respectively. |
| Meen (1944), Harding (1944) | Hastings and Grenville were separated, but both belonged to 1 group. Both were intruded by plutons. |
| Burns (1951), Ambrose and Burns (1956) | Distinction between the Hastings and Grenville considered unwarranted; introduced the term Flinton Group for conglomerates in the Flinton syncline. |
| Smith (1958) | Distinguished some sedimentary units, now defined as part of the Flinton Group, from older supracrustal rocks. Recognized unconformity between granite and conglomerate near Plevna Lake (now Sand Lake), but in Smith et al. (1969) suggested the contact was intrusive, not an unconformity. |
| Bain (1960), Lumbers (1964, 1967b), Hewitt (1964b, 1968), Hewitt and James (1956) | Supracrustal rocks assigned to the Grenville Series, divided into the Hermon and Mayo groups (and a variety of formations); some strata now assigned to the Flinton Group included in this scheme. Plutonic rocks considered to postdate all supracrustal rocks. |
| Wynne-Edwards (1967c) | First use of the term Grenville Group (instead of Series). |
| Wynne-Edwards (1972), Moore and Thompson (1972), Thompson (1972) | Grenville Supergroup established, Flinton Group noted as unconformable on the Mayo and Hermon groups but considered part of the Grenville Supergroup. Regional correlations presented in Wynne-Edwards (1972). Bishop Corners, Myer Cave, Fernleigh and Lessard formations established. Flinton Group postdates plutonic rocks. |
| Bright (1976, 1977, 1987a) | Established several new formations in the Harvey–Cardiff Arch which he correlated with the Hermon and Mayo groups. Established the Anstruther Lake Group, an arkosic sequence that underlies the Hermon Group. |
| Chappell (1978) | Established several stratigraphic terms in the Clare River area and correlated units with the Hermon and Flinton groups. |
| Moore and Thompson (1980) | Flinton Group formally defined with type sections. Flinton Group defined as postdating the Grenville Supergroup. Ore Chimney Formation (Grenville Supergroup) defined as a regolith underlying the Bishop Corners Formation. Correlations made with the Madoc area where the Madoc and Stewart formations are defined. |
| Bartlett (1983), Bartlett and Moore (1985) | Defined several formations in the Belmont Lake area for the supracrustal succession. |
| Abdel-Rahman and Martin (1987), Easton (1989a) | Suggested Flinton Group correlatives in the Madoc area older than the 1241 million-year-old Deloro Granite. Easton (1989a) suggested Madoc and Stewart formations are unaltered Grenville Supergroup rocks (as did Adams et al. 1907) and recommended abandonment of these units, as well as other disused stratigraphic terms in the Central Metasedimentary Belt. |
| Kinsman and Parrish (1990) | Provided maximum U-Pb age of 1157 ± 3 Ma on deposition of the Bishop Corners Formation meta-arenite confirming that the Flinton Group is considerably younger than the Grenville Supergroup. |
| Ford (in prep.), Easton and Ford (1991) | Refinement of the Myer Cave Formation stratigraphy in the Fernleigh area. Expanded Flinton Group to include other metaconglomerates and metapelites in the Fernleigh–Plevna area not previously included in the Flinton Group. Confirmed presence of unconformity and regolith below the Bishop Corners Formation. |

and metawacke, with interbedded mafic, commonly pillowed flows (Easton and Ford 1990a, 1990b, in prep.), and are termed, herein, the Grimsthorpe Group. Type areas are indicated in Table 19.13. Metaconglomerates in the lower part of the section contain gabbroic clasts similar to Canniff Complex and mafic volcanic fragments. The Grimsthorpe Group may be in unconformable contact with the Canniff Complex (Easton and Ford, in prep.) and is roughly correlative with the Tudor Formation of the Basal Group (*see* Table 19.7).

## MAZINAW TERRANE

### Mazinaw Group (New Term)

Supracrustal rocks in the central and northern Mazinaw Terrane are tentatively assigned to the Mazinaw Group, which can probably be further subdivided. Supracrustal rocks in the Clare River area are assigned to the Clare River Group, which may be roughly correlative with the Mazinaw Group. The Mazinaw Group includes both metavolcanic and metasedimentary rocks, including the Harlowe–Nervine Lake volcanic complex *of* Moore and Morton (1986) (Kashwakamak Formation *of* Harnois and Moore 1988), as well as voluminous marble and siliceous clastic metasedimentary rocks present in the Marble Lake–Plevna–Ompah area. Fault repetition of stratigraphy might be widespread in the northern Mazinaw Terrane (Easton and Ford 1991), and until the structural geology of the terrane is better understood, introduction of additional formations seems unwarranted.

### Kaladar Group

The supracrustal sequence, apart from Flinton Group correlatives, in the Clare River area constitutes the Kaladar Group (corresponding roughly to the Kaladar Group *of* Ambrose and Burns 1956). The subdivisions *of* Chappell (1978; *see* Table 19.14) are not easily correlated with Elzevir Terrane stratigraphic units, hence the exclusion from Table 19.7.

### Flinton Group

The Flinton Group is retained, much as defined by Moore and Thompson (1972, 1980; *see* Table 19.14), as is the Skootamatta Formation *of* Bain (1960). Chappell (1978) defined the Beatty and Bogart formations for the Flinton Group in the Clare River area because of correlation difficulties with the type area, although he did suggest that the Beatty and Bogart formations were roughly correlative with the Bishop Corners and Lessard formations, respectively. Bright (1986a) encountered similar correlation problems. Rocks in the Madoc area that Moore and Thompson (1980) correlated with the Flinton Group (Madoc and Stewart formations) are no longer considered to be Flinton Group correlatives (Abdel-Rahman and Martin 1987; Easton 1989a).

The only significant change in the stratigraphy of the Flinton Group is within the Myer Cave Formation (Ford, in prep.; Easton and Ford 1991; *see* Table 19.14). The black-sulphide schist of the Myer Cave Formation is subdivided into a lower, finely laminated, black schist overlain by a calcitic marble (50 to 250 m thick) or a marble megabreccia with a matrix of laminated, black schist. Overlying the calcite marble unit is an unlaminated black schist, with disseminated pyrite mineralization, that grades into the lowermost Fernleigh Formation.

The Flinton Group may be correlative with the Frontenac Supergroup, as it contains detritus of Frontenac Terrane rocks and has had a similar structural and metamorphic history. Flinton Group equivalents, however, have not yet been reported from the Frontenac Terrane.

## SHARBOT LAKE TERRANE

### Sharbot Lake Group (New Term)

The Sharbot Lake Group constitutes the supracrustal sequence within the Sharbot Lake Terrane. Three main sequences are present, with a marble succession predominating; each may in future be considered as formations. They are: 1) a lower, tholeiitic, mafic metavolcanic sequence, containing pillowed flows, tuff breccia and tuff, with minor intercalated black shales and interflow sediments; 2) calcite marbles, with dolomite marble interfingering with, and overlying, the top of the volcanic succession; and 3) a sequence of volcaniclastic metasedimentary rocks, including interlayered (bedded) para-amphibolite, amphibolite and metawackes which occur mainly along the southern and eastern margins of the domain. These metasedimentary rocks may be distal equivalents of the metavolcanic rocks and are probably slightly older than the carbonate sequence.

# Volcanism

## VOLCANIC ASSEMBLAGES

Volcanologic studies allow constraints to be placed on the nature and possible tectonic setting of volcanism. If volcanism in the Central Metasedimentary Belt occurred in an oceanic island-arc environment, as suggested by Brown et al. (1975), one might expect abundant pyroclastic rocks of intermediate to felsic composition (cf., Garcia 1978). If volcanism occurred in other environments, this should be reflected in the types of volcanic products, as well as in their chemistry.

Figure 19.57 outlines the distribution of volcanic and volcaniclastic rocks within the Central Metasedimentary Belt in Ontario. Figure 19.58 outlines the distribution of possible volcanic centres and the location of key environmental textures, such as ash-flow tuffs and pillowed flows. Figures 19.59a to 19.59f and Figures 19.60a to 19.60d illustrate the variety of volcanic textures preserved in Grenville Supergroup mafic and felsic metavolcanic rocks, respectively. Figures 19.60e and 19.60f and Figures 19.61a to 19.61d illustrate the variety of volcaniclastic metasedimentary rocks present within the Grenville Supergroup.

Many of the mafic metavolcanic rocks of the Central Metasedimentary Belt are characterized by extremely thin (less than 5 mm thick) pillow selvages, little hyaloclastite between pillows, only minor pillow breccia and few gas cavities or radial joints (*see* Figures 19.59a and 19.59b). Disseminated pyrrhotite is common in many mafic flows. These features suggest extrusion of volatile-poor, mafic lava, probably in deep water (greater than 300 m depth). These selvages are easily destroyed during deformation, and thus, the identification of pillowed sequences can be difficult. Pillowed sequences may be more widespread than illustrated in Figure 19.58. Only a few localities of ash-flow tuffs indicating subaerial deposition have been identified to date (*see* Figures 19.58 and 19.60b). In these areas, the presence of stromatolitic marbles closely associated with the volcanic sequences and abundant pyroclastic rocks (*see* Figure 19.60a) indicate broad areas of shallow-water deposition, thus subaerially erupted volcanic material may be more widespread than indicated in Figure 19.58.

Five main volcanic and volcanic-plutonic assemblages can be outlined across the Central Metasedimentary Belt (Table 19.15) on the basis of rock type, chemistry, related plutonic rocks, associated mineralization and geochronologic information. Assemblages $V_2$ and $V_3$ comprise much of the area of classic Grenville studies, and mafic rocks of these 2 assemblages have been designated in the past as the Tudor Formation of the Hermon Group. Much of the Tudor Formation described by Lumbers (1967a) and Sethuraman and Moore (1973) represents part of sequence $V_3$, corresponding to the Basal Group of Table 19.12. Volcanic units of the revised Hermon Group (*see* Table 19.12), such as the Oak Lake, Turriff, Crowe River and Burnt Lake formations (*see* Tables 19.12 and 19.15) correspond to assemblage $V_2$. These volcanic assemblages are separated mainly by major fault systems, commonly corresponding to internal terrane-domain boundaries and help define the internal stratigraphic-tectonic subdivisions present within the Elzevir Terrane. The only exception may be the assemblage $V_3$ to $V_4$ contact near Mazinaw Lake and in the Palmerston–Clyde–Calabogie (PCC) area, which could be a stratigraphic contact, although recent mapping suggests that this contact is also tectonic (Easton and Ford 1991).

Assemblage $V_2$ is dated at 1254 to 1248 Ma (Davis and Bartlett 1988). Later mafic plutonism occurred at 1242±3 Ma (Cordova Gabbro) contemporaneous with granitic plutonism at 1241±2 Ma (Deloro Granite, van Breemen and Davidson 1988b) and 1242±20 Ma (Methuen Granite, Heaman et al. 1986). Assemblage $V_3$ is cut by 1280 to 1270 million-year-old tonalites (Lumbers et al. 1990). Felsic rocks from the upper Tudor Formation have been dated at 1286±15 Ma (Silver and Lumbers 1966), and the Cordova Lake Formation in the Belmont Lake area is also likely part of assemblage $V_3$ and is $1287^{+11}_{-3}$ million years in age. Thus, roughly 30 my separates assemblage $V_2$ and $V_3$ in age. Easton and Ford (1990a, 1990b) report 2 volcanic sequences of different relative age within assemblage $V_3$ in the Grimsthorpe area. The younger sequence (Grimsthorpe Group) is chiefly mafic volcaniclastic sedimentary rocks derived from a tholeiitic volcanic source and related gabbroic intrusions, probably correlative with the Basal Group (*see* Table 19.7). The older sequence (Canniff Complex) consists of deformed tholeiitic flows and gabbroic rocks, which were deformed and hydrothermally altered prior to deposition of the volcaniclastic sediments. Thus, the Canniff Complex could range in age from 1285 Ma to over 1300 Ma.

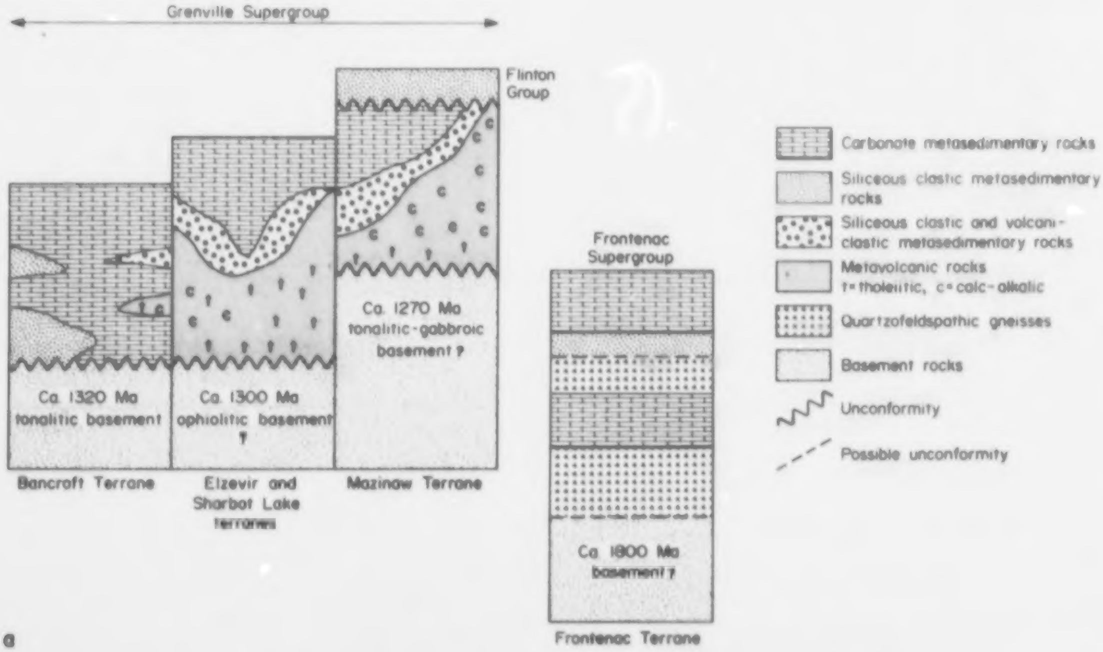

Grenville Supergroup
Flinton Group
Ca. 1320 Ma tonalitic basement
Ca. 1300 Ma ophiolitic basement ?
Ca. 1270 Ma tonalitic-gabbroic basement ?
Bancroft Terrane
Elzevir and Sharbot Lake terranes
Mazinaw Terrane
Frontenac Supergroup
Ca 1800 Ma basement ?
Frontenac Terrane
Carbonate metasedimentary rocks
Siliceous clastic metasedimentary rocks
Siliceous clastic and volcaniclastic metasedimentary rocks
Metavolcanic rocks t= tholeiitic, c= calc-alkalic
Quartzofeldspathic gneisses
Basement rocks
Unconformity
Possible unconformity
a

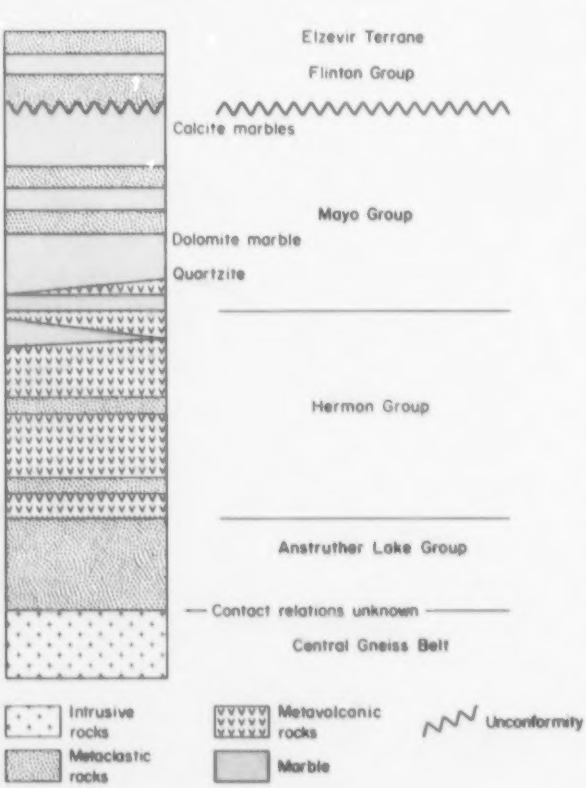
Figure 19.53a. Revised stratigraphic column for the Central Metasedimentary Belt in Ontario.

LeBaron et al. (1987) speculated that talcose rocks within assemblage $V_3$ (correlative with the Canniff Complex) adjacent to the Elzevir Tonalite represent altered komatiites, primarily on the basis of geochemistry. Di Prisco (1989) and Easton (1989b) concluded that the talcose rocks represent altered intrusive rocks (of roughly basaltic komatiite composition), rather than true komatiites.

Different types of mineralization are more closely associated with some assemblages than others (*see* Table 19.15). Assemblages $V_2$ and $V_4$ are more likely to host volcanogenic massive sulphide deposits, with $V_3$ and $V_5$ being more likely to host copper-nickel and, possibly, gold deposits.

## GEOCHEMISTRY

Two geochemical suites appear to be present in volcanic rocks of the Central Metasedimentary Belt, a more voluminous tholeiitic suite and a scarcer calc-alkalic suite. The tholeiitic suite may be of 2 ages, an older, *circa* 1290 million-year-old dominantly mafic succession (Grimsthorpe Domain, possibly the Sharbot Lake Terrane) and a younger, *circa* 1250 million-year-old bimodal subalkalic basalt-dacite-rhyolite suite (Harvey–Cardiff Arch, Belmont Domain). The calc-alkalic suite, possibly slightly younger in age, consists mainly of calc-alkalic basalt-andesite-dacite and rhyodacite and is only well-documented in 2 areas, the Stoney Lake area in the Belmont Domain and the Harlowe–Marble Lake area in the Mazinaw Terrane (Sethuraman and Moore 1973; Condie and Moore 1977; Moore and Morton 1986). Both calc-alkalic sequences are found on the western margins of the domains, near the domain boundaries. Bartlett (1983) reports calc-alkalic andesite intercalated within the lower tholeiitic basalt and upper subalkalic dacites of the Belmont area, suggesting that calc-alkalic and tholeiitic magmatism may have locally been contemporaneous. Details on the geochemistry of the various suites are presented below in the terrane descriptions.



b

Figure 19.53b. Generalized stratigraphic column for the Central Metasedimentary Belt in general use from 1967 to 1990 (*modified from* Lumbers 1967a, 1967b; Bright 1977; Carter 1984).

Various tectonic settings have been attributed to the bimodal tholeiite suite, including rifting of continental crust in a back-arc setting (Turriff metavolcanics, northern Belmont Domain, Holm et al. 1986) and ocean-floor basalts or an ensialic back-arc basin (Belmont Lake area, southern Belmont Domain, Bartlett 1983). Condie and Moore (1977) noted they were similar to low potassium Archean tholeiites and modern ocean and island-arc tholeiites (Figure 19.62) but did not assign a tectonic setting. The calc-alkalic sequence in the Harlowe–Marble Lake area has been attributed to an island-arc setting (Brown et al. 1975).

## VOLCANICLASTIC SEDIMENTATION

Many of the siliceous clastic metasedimentary rocks present within the Elzevir, Mazinaw and Sharbot Lake terranes are derived from volcanic sources (e.g., Figure 19.63). Some of these volcaniclastic metasedimentary rocks are geochemically indistinguishable from their volcanic sources (e.g., Apsley Formation, Figures 19.64 and 19.65). In many areas, such as in the Grimsthorpe Domain and in the Mazinaw Terrane, volcaniclastic metasedimentary rocks are volumetrically equal to, or more abundant than, the metavolcanic rocks from which they were likely derived. As shown in Figures 19.63 and 19.64, the volcaniclastic rocks probably represent debris aprons that developed around emergent volcanic islands.

## DEPOSITIONAL ENVIRONMENTS

Two main depositional environments appear to be represented in the Central Metasedimentary Belt as detailed above. The older, predominantly tholeiitic metavolcanic rocks appear to have been deposited in relatively deep water, probably, as mafic plains or shields in an oceanic setting. These mafic plains served as basement to the younger, more diverse, volcanic sequences.

The younger basalt-dacite tholeiitic sequences, and the few calc-alkalic sequences, were probably deposited mainly as volcanic islands, or as sea-facing volcanic edifices deposited along the edge of small landmasses. These volcanic islands were locally emergent, and a wide variety of sedimentary and volcanic environments are present at the land-sea interface, as illustrated in Figure 19.64. Offshore, turbidite fan systems served to transport clastic debris great distances from the volcanoes. These fans interfingered with carbonate turbidites being deposited within the same basin (*see* Figure 19.64).

Table 19.7. Stratigraphic framework for the Central Metasedimentary Belt used in this report.

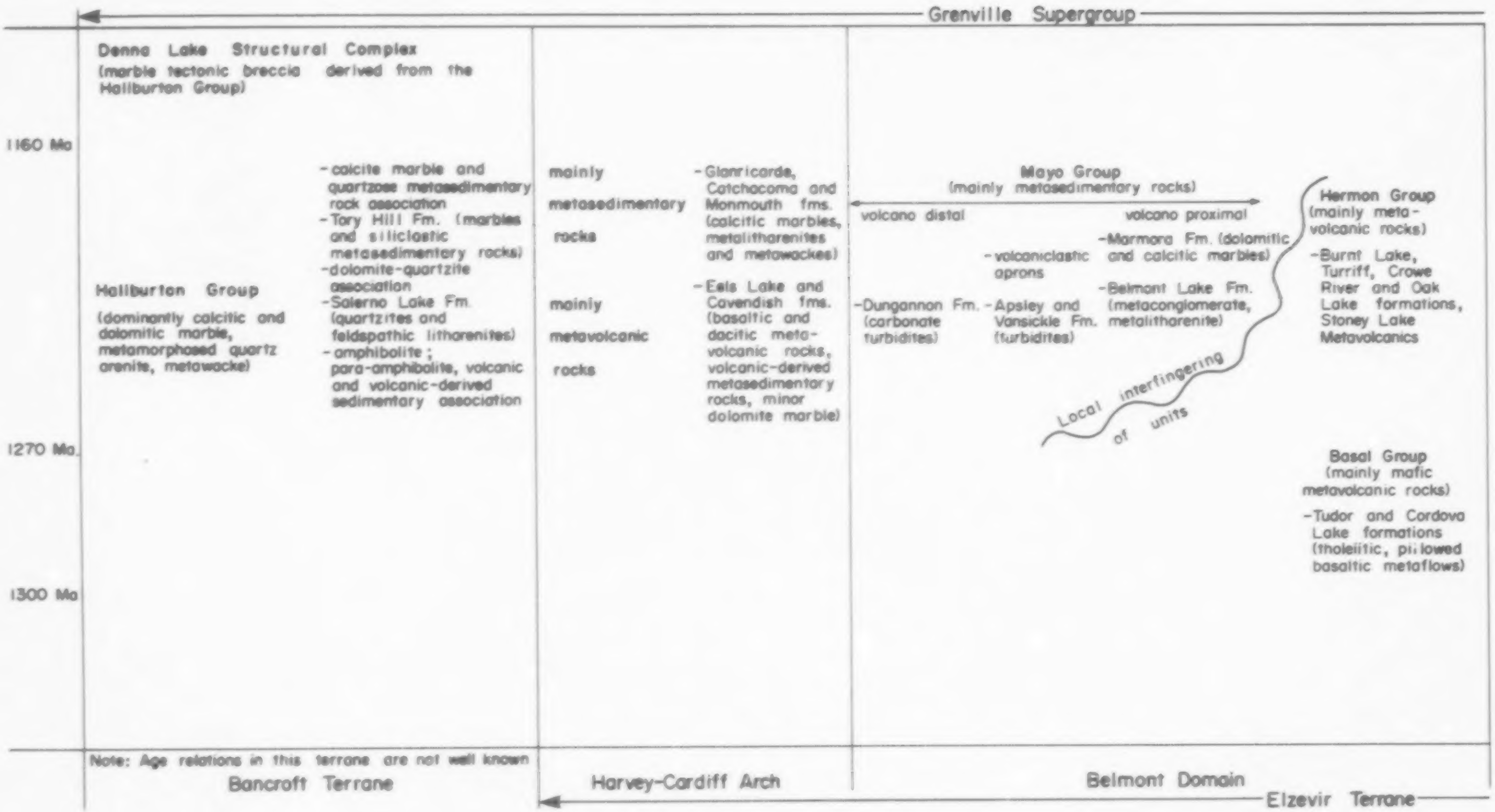

Grenville Supergroup
1160 Ma
1270 Ma
1300 Ma
Denna Lake Structural Complex
(marble tectonic breccia derived from the Haliburton Group)
Haliburton Group
(dominantly calcitic and dolomitic marble, metamorphosed quartz arenite, metawacke)
- calcite marble and quartzose metasedimentary rock association
- Tory Hill Fm. (marbles and siliciclastic metasedimentary rocks)
- dolomite-quartzite association
- Salerno Lake Fm. (quartzites and feldspathic litharenites)
- amphibolite; para-amphibolite, volcanic and volcanic-derived sedimentary association
Note: Age relations in this terrane are not well known
Bancroft Terrane
mainly metasedimentary rocks
mainly metavolcanic rocks
- Glanricarde, Catchacoma and Monmouth fms. (calcitic marbles, metalitharenites and metawackes)
- Eels Lake and Cavendish fms. (basaltic and dacitic metavolcanic rocks, volcanic-derived metasedimentary rocks, minor dolomite marble)
Harvey-Cardiff Arch
Mayo Group
(mainly metasedimentary rocks)
volcano distal
volcano proximal
- Dungannon Fm. (carbonate turbidites)
- volcaniclastic aprons
- Apsley and Vansickle Fm. (turbidites)
- Marmora Fm. (dolomitic and calcitic marbles)
- Belmont Lake Fm. (metaconglomerate, metalitharenite)
Local interfingering of units
Hermon Group
(mainly metavolcanic rocks)
- Burnt Lake, Turriff, Crowe River and Oak Lake formations, Stoney Lake Metavolcanics
Basal Group
(mainly mafic metavolcanic rocks)
- Tudor and Cordova Lake formations (tholeiitic, pillowed basaltic metaflows)
Belmont Domain
Elzevir Terrane

Table 19.7. Stratigraphic framework for the Central Metasedimentary Belt used in this report.

| | Grenville Supergroup | | | Frontenac Supergroup |
|---|---|---|---|---|
| 1160 Ma | | Flinton Group (metasedimentary rocks)<br>-Lessard Fm. (calcareous metasedimentary rocks, metaconglomerate)<br>-Fernleigh Fm. (calcareous turbidites)<br>-Myer Cove Fm. (black shales, calcitic marbles, graphitic schists)<br>-Bishop Corners Fm. (metaconglomerate, metamorphosed quartz arenite, metapelite) | | |
| | | unconformity<br>-Ore Chimney Fm. - schists (possible regolith) developed on Mazinaw Group supracrustal and metaplutonic rocks<br>Mazinaw Group<br>-calcitic and dolomite marbles, metawackes, volcanic-derived metasedimentary rocks<br>-Kashwakamak Fm. (basalt-andesite-dacite-rhyolite metavolcanic rocks)<br>Kaladar Group<br>-calcitic and dolomitic marbles, metawackes, volcanic-derived metasedimentary rocks<br>-basalt-andesite-dacite-rhyolite metavolcanic rocks<br>N ↔ S | Sharbot Lake Group<br>-calcitic marbles, minor dolomite marble and metawacke<br>-para-amphibolite, amphibolite, metawackes (mainly in southern part of terrane)<br>-tholeiitic, basaltic, pillowed, metavolcanic, minor volcanic-derived metasedimentary rock, minor dolomitic marble | |
| 1270 Ma | Grimsthorpe Group (mainly volcanic-derived metasedimentary rocks) - volcanic-derived metaconglomerate and metawacke, minor pillowed basalt metaflows, rusty schists<br>unconformity | | | -Upper Marble and Quartzite (calcitic marble, metamorphosed quartz arenite, minor metawacke)<br>-Main Gneiss Unit (pyroxene gneiss, quartzofeldspathic gneiss, para-amphibolite)<br>-Major Marble (calcite marble, calc-silicate gneiss)<br>-Lower Gneiss and Granulite (migmatitic quartzofeldspathic gneisses) |
| 1300 Ma | Conniff Complex - tholeiitic basalt pillowed metaflows, mylonitic metagabbro and ultramafic rocks | | | |
| | Grimsthorpe Domain<br>—Elzevir Terrane→ | Mazinaw Terrane | Sharbot Lake Terrane | Note: Age relations in this terrane are not well known<br>Frontenac Terrane |

**Table 19.8.** List of outdated and abandoned stratigraphic terms in the Central Metasedimentary Belt in Ontario.

| | |
|---|---|
| **Terms Suggested For Abandonment** | |
| Joe Lake Volcanics<br>(Sangster 1970; Carter 1981) | Term originally was poorly defined in terms of the extent of the unit, and differences from other stratigraphic units. Most of these rocks have subsequently been shown to be mylonitized gabbroic and mafic metavolcanic rocks (Easton 1988a). Term has no local significance and should be redefined or abandonded. |
| Madoc Volcanics<br>(Hewitt 1968) | As outlined in Easton (1989a), evidence for any significant amount of volcanic material within the Madoc Volcanics is minimal, although Abdel-Rahman and Martin (1990) do not agree. The unit consists dominantly of porcelainites formed in the alteration halo of the Deloro Granite (alteration is noted by both Easton 1989a and Abdel-Rahman and Martin 1990). Utility of the term is questionable and it should be redefined or abandonded. |
| Queensborough Acid Volcanics<br>(Hewitt 1968) | Most, if not all of these rocks are silicified metasedimentary rocks and felsites related to the alteration halo of the Deloro Granite (Easton 1989a; H.Meyn, MNDM Bancroft, Ontario, personnal communication, 1990). Term should be abandoned. |
| Stewart Formation<br>(Moore and Thompson 1980) | No indication that these rocks are distinctly different from other carbonate rocks in the immediate vicinity. Term has no local significance and should be abandonded. |
| Lasswade Formation<br>(Shaw 1962; Lumbers 1967a, 1967b) | No indication that these rocks are stratigraphically distinct from the Dungannon Formation (*see* Easton 1986d). Term should be abandoned. |
| **Terms Needing Redefinition** | |
| Madoc Formation<br>(Moore and Thompson 1980) | Term introduced for pelitic rocks thought to be correlative with the Flinton Group. These rocks have been variably altered by the Deloro Granite. Unit has local significance and should be retained. As there is no indication that the unit is younger than the Grenville Supergroup, it should be reassigned from the Flinton Group to the Grenville Supergroup. |
| Kaladar Complex<br>(Brown et al. 1975; Chappell 1978) | Kaladar Complex consists of rocks that now comprise the Canniff Complex, the Grimsthorpe Group and the Killer Creek Suite. Term needs to be redefined to take into account the diverse age range of the rock types composing the Kaladar Complex. Term should be redefined. |
| **Terms Redefined in this Paper** | |
| Anstruther Lake Group<br>(Bright 1976, 1977) | Redefined as the Anstruther Lake Lithodeme (Table 19.10). Lithodeme better reflects the gneissic character of the unit and the problematic origin of the unit (meta-arkose *versus* meta-plutonic rock). |
| Hermon and Mayo groups | Redefined in Table 19.12 to include additional formations proposed since initial establishment. |

**Table 19.9.** Stratigraphic usage in the Bancroft Terrane. Units listed in approximate stratigraphic order from oldest to youngest.

| | |
|---|---|
| Denna Lake Structural Complex<br>(Easton 1983, 1985b, 1987a, 1987e, 1990a) | The Denna Lake Structural Complex consists predominantly of several types of marble tectonic breccia that can be mapped regionally on the basis of dominant clast type within the breccias. All rock units of the Haliburton Group are present within the Denna Lake Structural Complex, as are dismembered metaplutonic rocks (mainly gneissic granites and meta-anorthosites). |
| **Haliburton Group*** | |
| Interbedded calcite marble and quartzose sedimentary rock association | Thinly interbedded calcitic marbles, silicated marble, calcareous metawackes and feldspathic metalitharenite. In the Tory Hill area, rocks of this association were assigned by Bright (1976, 1977) to the Tory Hill Formation. |
| | The Tory Hill Formation consists of thinly interbedded limestone, calcareous greywacke, arkosic sandstone, chert and subordinate tuff (calcitic silicated marble, calc-silicate gneiss, quartzofeldspathic gneiss, impure quartzite, minor amphibolite) (<30 to 1350 m).Thins rapidly westward into the Howland area (Easton 1987a). |
| Dolomite-quartzite association | Dolomitic marble, metamorphosed quartz arenites and feldspathic metalitharenites. In the Eels Lake area, the siliclastic rocks have been separated into the Salerno Lake Formation by Bright (1976, 1977). |
| | The Salerno Lake Formation consists of arkosic sandstone and thick quartz arenite units (biotite-poor quartzofeldspathic gneiss and fine- to medium-grained quartzite) (240 to 600 m). Mainly northwest Cavendish and southwest Glamorgan townships. |
| Amphibolite, volcanic and volcaniclastic association | Mafic to intermediate tuffs and lapilli tuffs, para-amphibolite and volcanic-derived metawackes (distally redeposited volcanic facies regime), subordinate metasiltstone and metalitharenite. Minor orthoamphibolite. |

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*Note: The Haliburton Group contains several lithologic associations that have not been defined as formal stratigraphic units, however, these lithologic associations could serve as the basis for future stratigraphic subdivision of the Haliburton Group. Many rock types in the Haliburton Group have been chemically modified by a *circa* 1050 million-year-old alkali metasomatic event producing "fenite" or "skarn-like" alteration assemblages.

**Table 19.10.** Stratigraphic usage in the Harvey–Cardiff Arch, Elzevir Terrane (*modified from* Bright 1976, 1977). Units listed in approximate stratigraphic order from oldest to youngest.

| Possible Hermon Group Correlatives | |
|---|---|
| Eels Lake Formation, upper clastic member | Thinly interbedded calcareous and feldspathic greywacke; subordinate tuff, marble and quartzite (calc-silicate gneiss, biotitic quartzofeldspathic gneiss, impure quartzite, minor amphibolite) (150 to 300 m). Regionally distributed, thinnest in Tory Hill–Gooderham area. May be correlative with parts of the Haliburton Group. |
| Eels Lake Formation, lower volcanic member | Mafic to intermediate volcanic rocks, minor felsic volcanic and sedimentary rocks (amphibolite, calc-silicate gneiss, feldspathic and aluminous gneiss) (300 to 900 m). Thickest sequences occur in southwest Cardiff Township and southeast Anstruther Township. |
| Cavendish Formation | Mafic to intermediate volcanic rocks, subordinate felsic volcanic rocks, minor clastic sedimentary rocks and marble (amphibolite, biotitic quartzofeldspathic gneiss and aluminous gneiss) (600 to 1500 m). Thickest in southwest Cavendish Township; probably stratigraphic equivalent of the lower volcanic member of the Eels Lake Formation. |
| Glanricarde Formation | Arkosic sandstones and siltstones, subordinate feldspathic and calcareous greywacke (biotite-poor quartzofeldspathic gneiss, subordinate biotite-rich gneiss, calc-silicate gneiss) (150 to 450 m). Mainly occurs in northern Anstruther Township. |
| Catchacoma Formation (originally termed Mississagua Formation in Bright 1976) | Interbedded feldspathic arenite and greywacke, subordinate calcareous greywacke, minor tuff (biotite- and hornblende-biotite quartzofeldspathic gneiss, subordinate calc-silicate gneiss and amphibolite) (<150 to 600 m). Thickest in southern Cavendish Township. |
| Monmouth Formation | Limestone and dolostone, minor greywacke, siltstone (calcitic to dolomitic marble, minor biotite-rich gneiss) (<30 to 900 m). Occurs mainly in the eastern half of the Eels Lake area (NTS 31 D/16). May be stratigraphically equivalent to the Dungannon Formation (Mayo Group?) |
| Anstruther Lake Lithodeme, upper subunit | Biotite-poor to biotite-rich quartzofeldspathic gneiss, minor calc-silicate gneiss (300 to 1500 m). Mainly present in southeastern part of the Eels Lake area (NTS 31 D/16). |
| Anstruther Lake Lithodeme, lower subunit | Biotite-poor quartzofeldspathic gneiss ±1 to 5% magnetite (300 to 1500 m). Mainly present in northwestern part of the Eels Lake area (NTS 31 D/16). Bright (1976, 1977) regarded this unit as a metamorphosed sequence of arkoses and feldspathic litharenites. In the Minden area, Easton (1987a, 1987e, 1990a) considered lithologically similar rocks rimming the Glamorgan Gneiss Complex as gneissic granites and classified them as metaplutonic rocks (Crego Lake Lithodeme of the Glamorgan Gneiss Complex). Because of the uncertainity in interpretation, the Anstruther Lake Group *of* Bright (1976, 1977) is redefined herein as the Anstruther Lake Lithodeme, as described above. This redefinition allows retention of this unit as a distinctive map unit without having to resolve the nature of the protolith of the unit. |

**Table 19.11.** Hermon Group and Mayo Group stratigraphy as outlined by Lumbers (1967a, 1967b, 1969)

| Grenville Supergroup | |
|---|---|
| **Mayo Group** | |
| Lasswade Formation | Marble, minor calcareous metasandstone and metasiltstone, rare recrystallized chert. |
| Apsley Formation | Poorly sorted feldspathic metasandstone with upper and lower members of calcareous metasandstone and metasiltstone. |
| Dungannon Formation | 80% marble, predominantly calcitic; remainder mainly calcareous metasandstone and metasiltstone, poorly sorted feldspathic metasandstone, well-sorted quartz-rich metasandstone and rare recrystallized chert; rare mafic metavolcanic flows and iron formation near base of formation overlying Tudor metavolcanics. |
| **Hermon Group** | |
| Burnt Lake Formation | Mainly rhyolite, trachytic, dacitic and andesitic metavolcanic rocks with minor metabasaltic flows; lensoid marble units and sandy siliceous metasedimentary rocks found locally. |
| Turriff Formation | 75% pillowed basaltic and andesitic metavolcanic flows; remainder dacitic and felsic flows and pyroclastic rocks, minor iron formation. |
| Vansickle Formation | Mainly poorly sorted feldspathic metasandstone with abundant marble, felsic and mafic metavolcanic rocks, metaconglomerate and well-sorted quartz-rich metasandstone; arkosic metasedimentary rocks commonly associated with felsic metavolcanic rocks. |
| Oak Lake Formation | Mainly metamorphosed felsic pyroclastic rocks and arkose with some mafic and felsic metavolcanic flows; rare marble and metaconglomerate in upper part of formation. |
| Tudor Formation | Mainly metamorphosed, pillowed, basaltic and andesitic flows; minor dacitic and felsic metavolcanic rocks and rare iron formation. |

**Table 19.12.** Stratigraphic usage in the Belmont Domain, Elzevir Terrane.

| **Grenville Supergroup** | |
|---|---|
| **Mayo Group** | |
| Marmora Formation (Bartlett 1983; deKemp 1984; Bartlett and Moore 1985) | Mainly dolomitic (commonly stromatolitic) and calcitic marbles, with interbedded calcareous metasiltstone and metasandstone, minor rusty schist. Represents a shallow water to ramp carbonate facies (deKemp 1984; Bourque 1982) and is distinct from the Dungannon Formation (although it could be a laterally facies equivalent of the Dungannon Formation). |
| Belmont Lake Formation (Bartlett 1983; Bartlett and Moore 1985) | Mainly metaconglomerate, metalitharenite and feldspathic litharenite, dolomitic marble, commonly stromatolitic, minor metasiltstone and metamudstone. Most sedimentary units are volcanic-derived and from the adjacent Belmont Lake Metavolcanic Complex. |
| Dungannon Formation (Hewitt and James 1956; Lumbers 1967b) | 80% marble, predominantly calcitic; remainder mainly calcareous metasandstone and metasiltstone, poorly sorted feldspathic metasandstone, well-sorted quartz-rich metasandstone and rare recrystallized chert. |
| Apsley Formation (Simony 1960; Shaw 1962; Lumbers 1967b; Easton 1986d, 1987f, 1989e) | Poorly sorted, volcanic-derived dacitic composition metasandstone with lower member of volcanic-derived, andesitic composition, metasandstone and metasiltstone and distal members of calcareous metasandstone and metasiltstone which interfinger with the Dungannon Formation. The Apsley Formation was probably deposited as a turbidite fan (Easton 1986d, 1987f, 1989e). |
| Vansickle Formation (Lumbers 1967b) | Mainly poorly sorted feldspathic metasandstone with abundant marble, felsic and mafic metavolcanic rocks, metaconglomerate and well-sorted quartz-rich metasandstone; arkosic metasedimentary rocks commonly associated with felsic metavolcanic rocks. Vansickle Formation was probably deposited as a turbidite fan and is probably roughly correlative with the Apsley Formation. Vansickle Formation probably interfingers with the Dungannon Formation distal from the volcanic centres. |
| **Hermon Group** | |
| Burnt Lake Formation (Lumbers 1967b) | Mainly rhyolite, trachytic, dacitic and andesitic metavolcanic rocks with minor metabasaltic flows; lensoid marble units and sandy siliceous metasedimentary rocks found locally. Probably *circa* 1260 to 1250 million years in age. Tholeiitic affinities. |
| Turriff Formation (Hewitt and James 1956; Lumbers 1967b) | 75% pillowed basaltic and andesitic metavolcanic flows; remainder dacitic and felsic flows and pyroclastic rocks, minor iron formation. Probably *circa* 1260 to 1250 million years in age. Tholeiitic affinities. |
| Crowe River Formation (Bartlett 1983; Bartlett and Moore 1985) | Mainly basaltic to andesitic metavolcanic flows, some of which were deposited subaerially. Probably *circa* 1250 million years in age. Composes part of the Belmont Lake Metavolcanic Complex. |
| Oak Lake Formation (Lumbers 1967b; Bartlett 1983; Bartlett and Moore 1985) | Mainly metamorphosed felsic pyroclastic rocks and arkose with some mafic and felsic metavolcanic flows; rare marble and metaconglomerate in upper part of formation. Probably *circa* 1250 million years in age. Tholeiitic affinities. Composes part of the Belmont Lake Metavolcanic Complex. |
| **Basal Group (no formal name yet proposed)** | |
| Cordova Lake Formation (Bartlett 1983; Bartlett and Moore 1985) | Mainly metamorphosed, pillowed basaltic to andesitic flows; minor dacitic metavolcanic rocks. Probably *circa* 1290 to 1280 million years in age. Tholeiitic. Composes part of the Belmont Lake Metavolcanic Complex. |
| Tudor Formation (Lumbers 1967b) | Mainly metamorphosed, pillowed, basaltic and andesitic flows; minor dacitic and felsic metavolcanic rocks and rare iron formation. Probably *circa* 1290 to 1280 million years in age. Tholeiitic. |

## Carbonate and Siliciclastic Sedimentation

Our knowledge of the distribution of sedimentary environments, particularly with respect to the carbonate rocks, is still embryonic. However, 2 broad subdivisions can be made, which correspond roughly with the terrane boundaries in the Central Metasedimentary Belt. Furthermore, within the Elzevir Terrane, internal domain boundaries serve to separate largely sedimentary (mainly carbonate) "basins" from "volcanic piles." Figures 19.66a to 19.66f and 19.67a to 19.67f illustrate the variety of metasedimentary rocks present within the Central Metasedimentary Belt.

## BIOSEDIMENTARY STRUCTURES

The recognition of fossilized life-forms in Grenville carbonate rocks dates back to 1865, when Dawson proclaimed an organic origin for the *Eozoon canadense* forms found by Sir William Logan in 1858 (Wilson 1957). *E. canadense* consisted of banded calcium carbonate and serpentine or tremolite, which Osann (1902) attributed to 2 periods of alteration. *E. canadense* comprises several similar, but distinct, morphologies (cf., Logan 1867). Recently, a biogenic origin for one of these forms, the *huntingdon* type (Dawson 1865, 1875, 1888; Logan 1867; Hofmann 1982) (named for the type locality at the Henderson talc mine in Huntingdon Township near Madoc), has gained general acceptance, mainly, due to the work of

**Table 19.13.** Stratigraphic usage in the Grimsthorpe Domain, Elzevir Terrane. Units listed in approximate stratigraphic order from oldest to youngest.

| | |
|---|---|
| Grimsthorpe Group* (new term) | Metasedimentary rocks derived from mafic volcanic and gabbroic source region and minor metabasalt flows. Matrix- and clast-supported metaconglomerate, gabbro cobble metaconglomerate, meta-arenite, metawacke, minor metasiltstone, metapelite and rusty schists. Minor pillowed and massive tholeiitic metabasalt flows. Named after Grimsthorpe Township, where the unit is well-exposed at low metamorphic grade and where most lithologic types within the unit are exposed. Type areas are: 1) the islands in the east-central part of Lingham Lake, 2) along the forest access road located between the Skootamatta Road and Lingham Lake and 3) along the Skootamatta Road between Killer Creek and Partridge Creek. |
| | *unconformity?* |
| Canniff Complex (new term) | Massive and pillowed tholeiitic metabasalts, spilitic metabasalts; protomylonitic metagabbro and metaperidotite. Ultramafic rocks commonly altered to talc ± anthophyllite ± actinolite rocks. (Note: Kaladar Complex *of* Chappell [1978] included elements of the Canniff Complex, Grimsthorpe Group and Killer Creek Suite metagabbros—detailed subdivision into component elements is preferred to the use of the term Kaladar Complex.) Named after Canniff Creek along which gabbroic members of the unit are well-exposed. Type areas are: 1) the area east of the Canniff Tonalite along Canniff Creek, 2) the area between the southern lobe of the Weslemkoon Tonalite and the Canniff Tonalite and 3) the northeast shore of Lingham Lake. |

*Grimsthorpe Group intruded by *circa* 1270 million-year-old and older plutons.

**Table 19.14.** Flinton Group stratigraphy (*modified from* Bain 1960; Chappell 1978; Moore and Thompson 1980; Ford, in prep.; Easton and Ford 1991). Type and reference localities listed in Moore and Thompson (1980).

| SW | | | NE |
|---|---|---|---|
| **Tweed** | **Clare River Area*** | **Flinton** | **Fernleigh and/or Ardoch** |
| | **Bogart formation**<br>– calcareous psammite<br>– dolomite, rusty pelite<br>– calcareous metawacke<br>– calcareous pelite<br>– calcite marble | **Lessard Formation**<br>– biotite-carbonate schist, thin-bedded<br>– calcareous quartzo-feldspathic psammite<br>– calc-silicate paragneiss<br>– plutonic cobble conglomerate | **Fernleigh Formation**<br>– biotite-carbonate schist, thin- to medium-bedded<br>**Myer Cave Formation**<br>– graphitic, rusty pelite, unlaminated<br>– graphitic, rusty pelite, laminated<br>– calcite marbles, carbonate conglomerate and megabreccia, pelite<br>– graphitic, rusty pelite, laminated, and biotite-staurolite-garnet schist<br>– dolomite marble<br>– biotite-staurolite-garnet schist |
| **Skootamatta Formation**<br>– volcanic pebble conglomerate<br>– carbonate pebble conglomerate<br>– calcareous phyllite, psammite | **Beatty formation**<br>– pelitic schist<br>– rusty psammite | **Bishop Corners Formation**<br>– micaceous pelitic schist, thinly bedded<br>– quartz arenite pebble conglomerate<br>– plutonic cobble conglomerate<br>– quartz arenite<br>– white quartz arenite cobble conglomerate | |

*informal units of Chappell 1978

Marika Bourque (1981, 1982; Bourque et al. 1982) and others (deKemp 1984). *Eozoon canadense huntingdon* is an algal-laminite stromatolite. Other *E. canadense* types, however, may not be biosedimentary in origin and should be regarded for the moment as pseudofossils (Hofmann 1971, 1982).

In addition to algal-laminite stromatolites, 5 types of stromatolite morphology have been documented from the Ontario Grenville, as shown in Figure 19.68. Type localities for all 5 types (*see* Figures 19.67c to 19.67f) are located in the western Elzevir Domain, mainly, in the Madoc area (Bartlett and deKemp 1987). They typically occur as laminited silicified forms in siliceous dolomitic marble (*see* Figures 19.67c and 19.67f); only rarely are they hosted in calcitic marble. deKemp (1984) suggests that dolomitization is for the most part a diagenetic process, much as occurs in modern environments, but that direct precipitation of dolomite may have only occurred locally. Silicification also seems to play an important role in preservation (deKemp 1984).

Stromatolites are useful in paleoenvironmental reconstructions, however, care should be taken when comparing more recent environments with Precambrian environments. The organisms responsible for the stromatolites were probably blue-green algae and bacteria, and, thereby, restricted to the photic zone (water depth less than 200 m). Playford et al. (1976) suggest 100 m was the maximum depth for Precambrian stromatolite growth. The absence of grazing invertebrates and vertebrates in the Precambrian would have made stromatolite growth possible in a wider variety of environments. Variations in stromatolite morphology are determined not only by water depth, but by prevailing energy systems, by growth mechanisms and by the biotic processes of the algae. In general, mats form in lower energy environments, and domal and columnar forms, form in higher stress regimes.

It should also be noted that a variety of carbonate environments are present within the Central Metasedimentary Belt, as outlined below (cf., Figure 19.69), and that the localized presence of stromatolites within carbonate belts in the Central Metasedimentary Belt do not point to a uniform shallow-water, carbonate depositional setting across the Central Metasedimentary Belt.

## PLATFORM-SLOPE ENVIRONMENTS

Platform-slope environments in the Bancroft and Frontenac terranes are characterized by abundant quartz arenites and thick sequences of dolomitic and calcitic marbles. Both terranes are preserved at high metamorphic grade, so detailed facies analysis is not possible.

In the Bancroft Terrane, regional changes in facies distribution suggests that the sedimentary basin deepened to the southeast. Volcaniclastic metasedimentary rocks and distally deposited metavolcanic rocks become more common in the southeast and are probably derived from sources that lay farther offshore (*see* Terrane Descriptions for details).

In the Frontenac Terrane, it appears that the basin deepened to the west and northwest, with deeper water facies being located in the area between the Rideau Lake fault and the Sharbot Lake Terrane boundary. Shallower water environments predominate along the St. Lawrence River, and across the river in upper New York State, evaporitic environments may be present (*see* Terrane Descriptions for details).

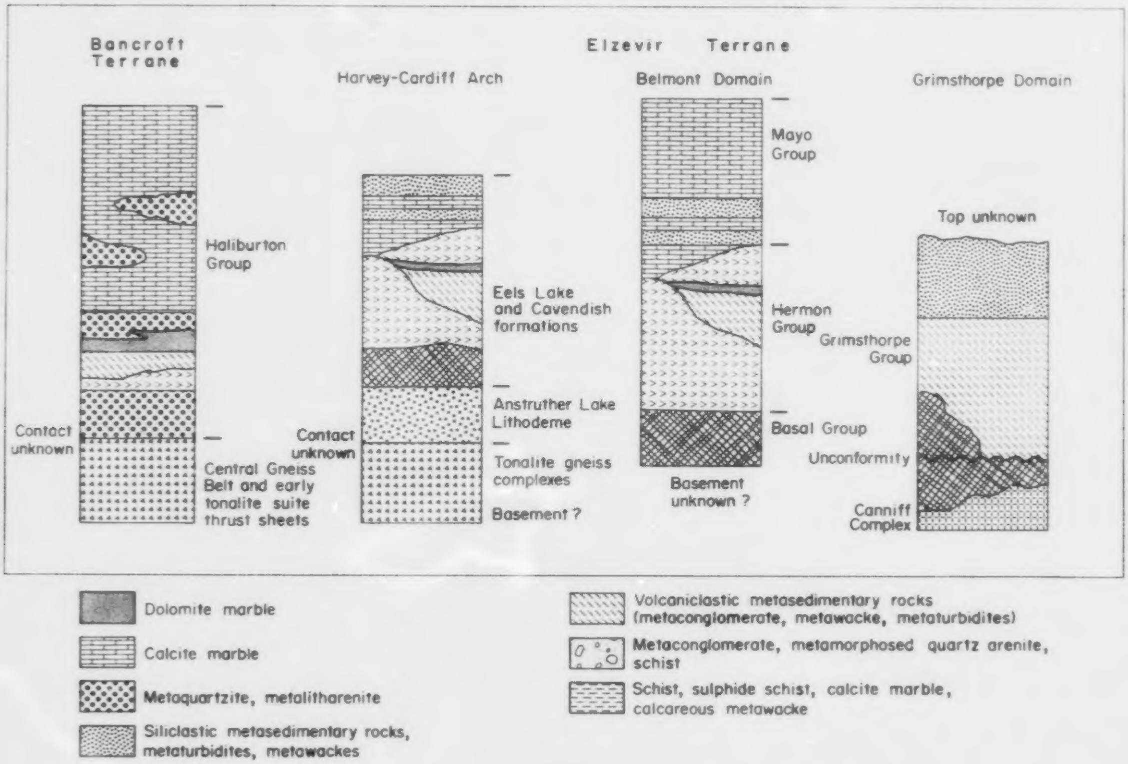

Bancroft Terrane
Harvey-Cardiff Arch
Elzevir Terrane
Belmont Domain
Grimsthorpe Domain
Haliburton Group
Contact unknown
Central Gneiss Belt and early tonalite suite thrust sheets
Contact unknown
Eels Lake and Cavendish formations
Anstruther Lake Lithodeme
Tonalite gneiss complexes
Basement ?
Mayo Group
Hermon Group
Basal Group
Basement unknown ?
Top unknown
Grimsthorpe Group
Unconformity
Canniff Complex
Dolomite marble
Calcite marble
Metaquartzite, metalitharenite
Siliclastic metasedimentary rocks, metaturbidites, metawackes
Volcaniclastic metasedimentary rocks (metaconglomerate, metawacke, metaturbidites)
Metaconglomerate, metamorphosed quartz arenite, schist
Schist, sulphide schist, calcite marble, calcareous metawacke

**Figure 19.54.** Very generalized stratigraphic sections across the Central Metasedimentary Belt showing major lithologic differences between domains and terranes.

## VOLCANIC ISLAND-SLOPE-PLAIN ENVIRONMENTS

Little is known about the relationship of carbonate sedimentation in environments undergoing active volcanism and volcaniclastic sedimentation. The Central Metasedimentary Belt, despite the presence of greenschist facies and locally higher metamorphic conditions, provides an opportunity to study these relationships.

Preliminary work (Easton 1987e, 1987g, 1989f, 1989g) indicates that several carbonate depositional environments are associated with Grenville volcanic complexes (probably volcanic islands; Figure 19.70), and include:

1. In the proximal volcanic environment, stromatolitic dolomite reefs develop adjacent to the island volcanoes (e.g., eastern Dungannon Formation). These dolomite reefs and associated marly and black shales serve as favourable hosts for zinc mineralization.
2. In the proximal to distal volcanic environment, deposition of carbonate turbidites derived from an adjacent carbonate platform occurred basinward of the volcanic islands. The carbonate turbidites interfinger with, or are buried by, prograding turbidite fans of both volcanic (Apsley Formation) and sedimentary (Vansickle Formation) provenance. Rhythmically layered couplets of ash and calc-silicate or marble with 2 to 10 cm scale bedding are common in the turbidite fans of pyroclastic-volcaniclastic origin and indicate simultaneous carbonate sedimentation and volcanic ash deposition.
3. In the distal volcanic environment, carbonate deposition occurred as turbidites in deep basins and as shelf and platform deposits in shallower areas.

## Magmatism in the Central Metasedimentary Belt

Nine major plutonic suites are documented within the Central Metasedimentary Belt, as summarized in Table 19.16. The distribution of these suites is illustrated in Figures 19.71a to 19.71e. The classification used here is based on that of Lumbers et al. (1990), the difference being slight changes in terminology. Most of our knowledge of plutonic rocks in the Central Metasedimentary Belt comes from the pioneering work of Lumbers (Lumbers 1967a, 1967b, 1982; Schwerdtner and Lumbers 1980). Other syntheses, mainly in the Central Metasedimentary Belt, include Sauerbrei (1966), Wynne-Edwards (1967b) and Britton (1979). Detailed work on specific plutons is outlined in the detailed terrane descriptions. An extensive body of work on the $^{18}O$ isotope characteristics of Grenville plutonic rocks and gneisses is found in Shieh (1985), Shieh and Schwarz (1974, 1978) and Shieh et al. (1976). Representative analyses of most suites are presented in Lumbers et al. (1990).

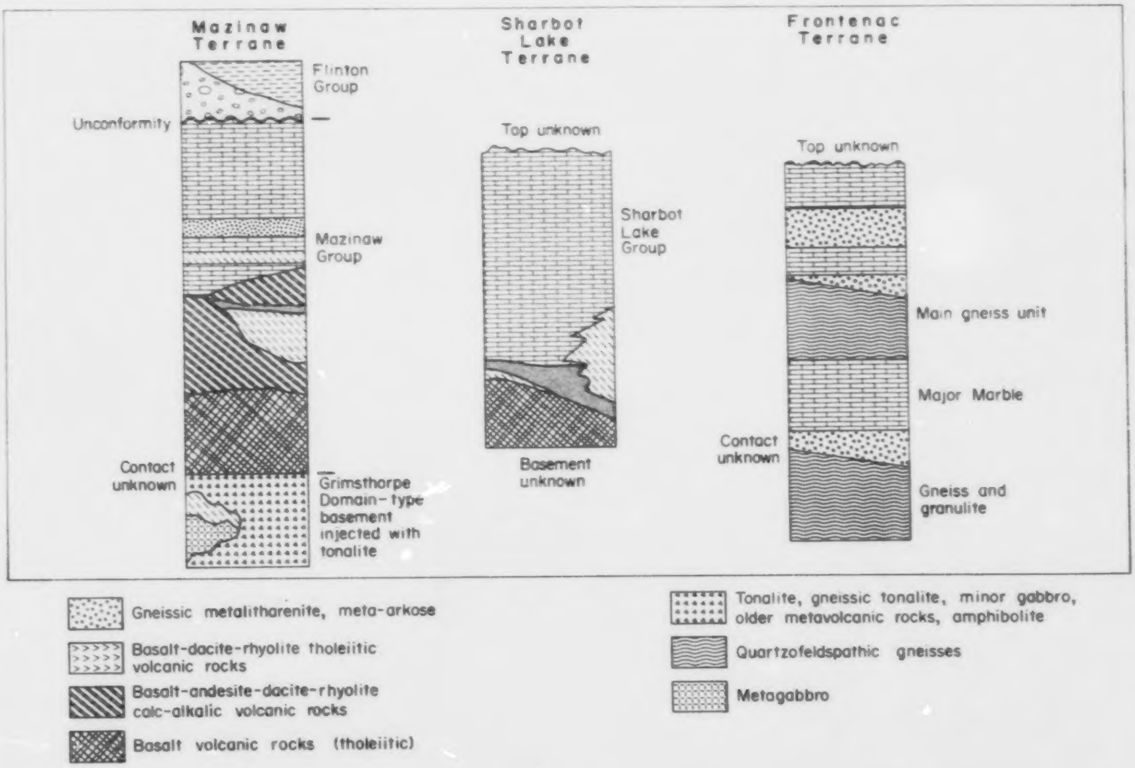
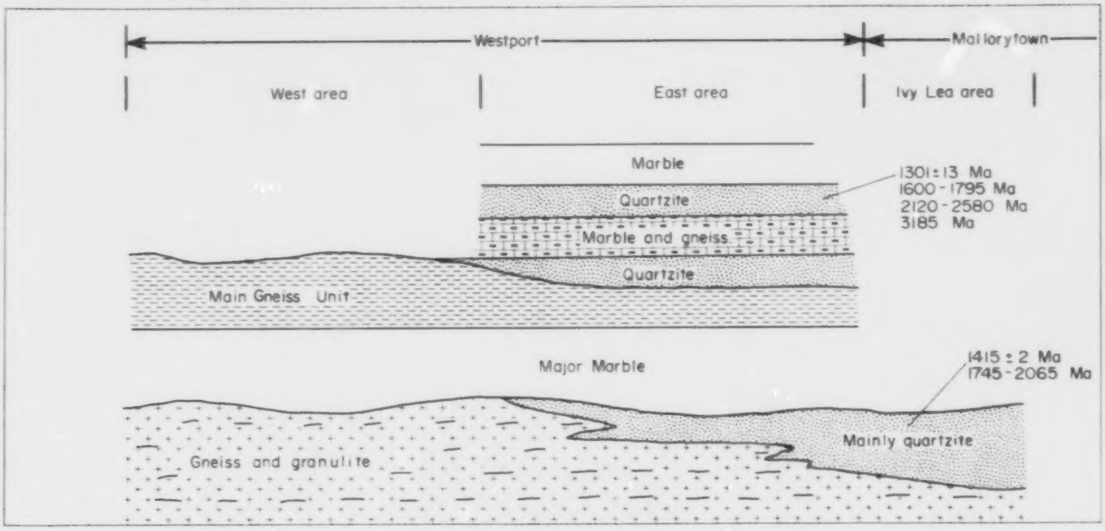

Mazinaw Terrane
Flinton Group
Unconformity
Mazinaw Group
Contact unknown
Grimsthorpe Domain-type basement injected with tonalite
Sharbot Lake Terrane
Top unknown
Sharbot Lake Group
Basement unknown
Frontenac Terrane
Top unknown
Main gneiss unit
Major Marble
Contact unknown
Gneiss and granulite
Gneissic metalitharenite, meta-arkose
Basalt-dacite-rhyolite tholeiitic volcanic rocks
Basalt-andesite-dacite-rhyolite calc-alkalic volcanic rocks
Basalt volcanic rocks (tholeiitic)
Tonalite, gneissic tonalite, minor gabbro, older metavolcanic rocks, amphibolite
Quartzofeldspathic gneisses
Metagabbro

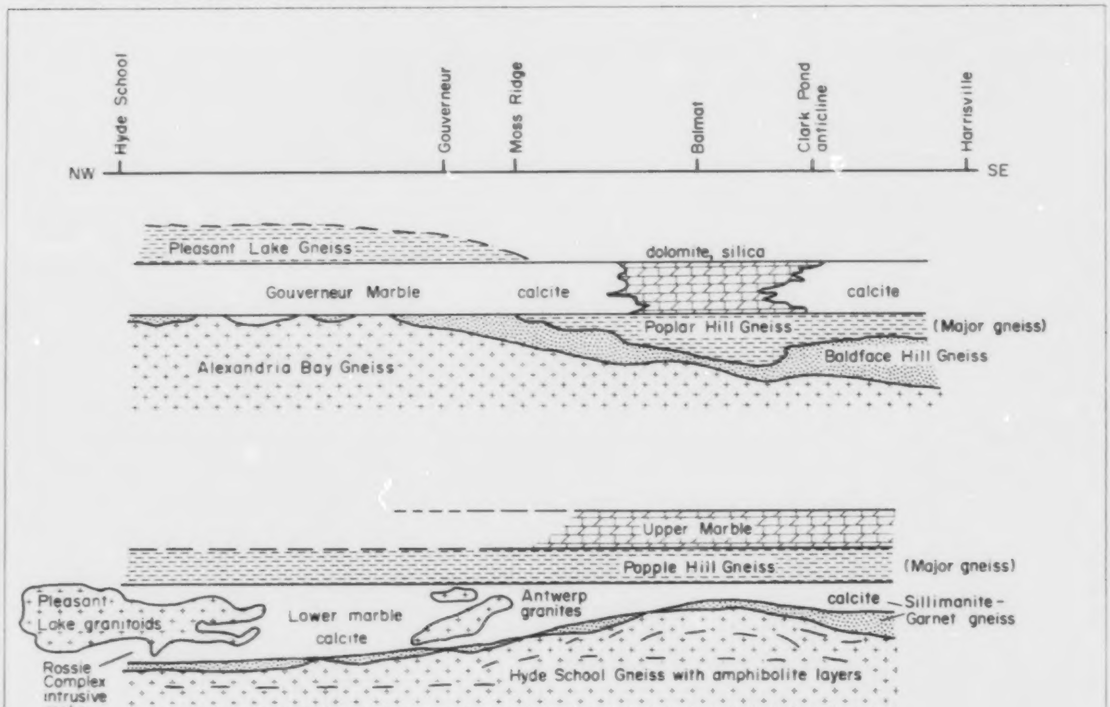
**Figure 19.54.** Very generalized stratigraphic sections across the Central Metasedimentary Belt showing major lithologic differences between domains and terranes.

Frontenac Supergroup: southeastern Ontario


Westport
Mallorytown
West area
East area
Ivy Lea area
Marble
Quartzite
Marble and gneiss
Quartzite
Main Gneiss Unit
1301±13 Ma
1600 - 1795 Ma
2120 - 2580 Ma
3185 Ma
Major Marble
1415 ± 2 Ma
1745 - 2065 Ma
Mainly quartzite
Gneiss and granulite

a

Frontenac Supergroup: northwestern Adirondacks


Hyde School
Gouverneur
Moss Ridge
Balmat
Clark Pond anticline
Harrisville
NW
SE
Pleasant Lake Gneiss
dolomite, silica
Gouverneur Marble
calcite
calcite
Poplar Hill Gneiss
(Major gneiss)
Alexandria Bay Gneiss
Baldface Hill Gneiss
Upper Marble
Popple Hill Gneiss
(Major gneiss)
Pleasant Lake granitoids
Lower marble calcite
Antwerp granites
calcite
Sillimanite - Garnet gneiss
Rossie Complex intrusive rocks
Hyde School Gneiss with amphibolite layers

b

**Figure 19.55. a)** Generalized stratigraphic relationships within the Frontenac Terrane (*modified from* Wynne-Edwards 1967a and Carmichael et al. 1987). U-Pb data of Kinsman (1990). **b)** Generalized stratigraphic relationships within the Adirondack Lowlands (*modified from* Wiener et al. 1984 [upper] and deLorraine and Carl 1986 [lower]). If assumptions made by Wynne-Edwards (1967a) are incorrect, section proposed by deLorraine and Carl (1986) may better reflect Frontenac Terrane stratigraphy.

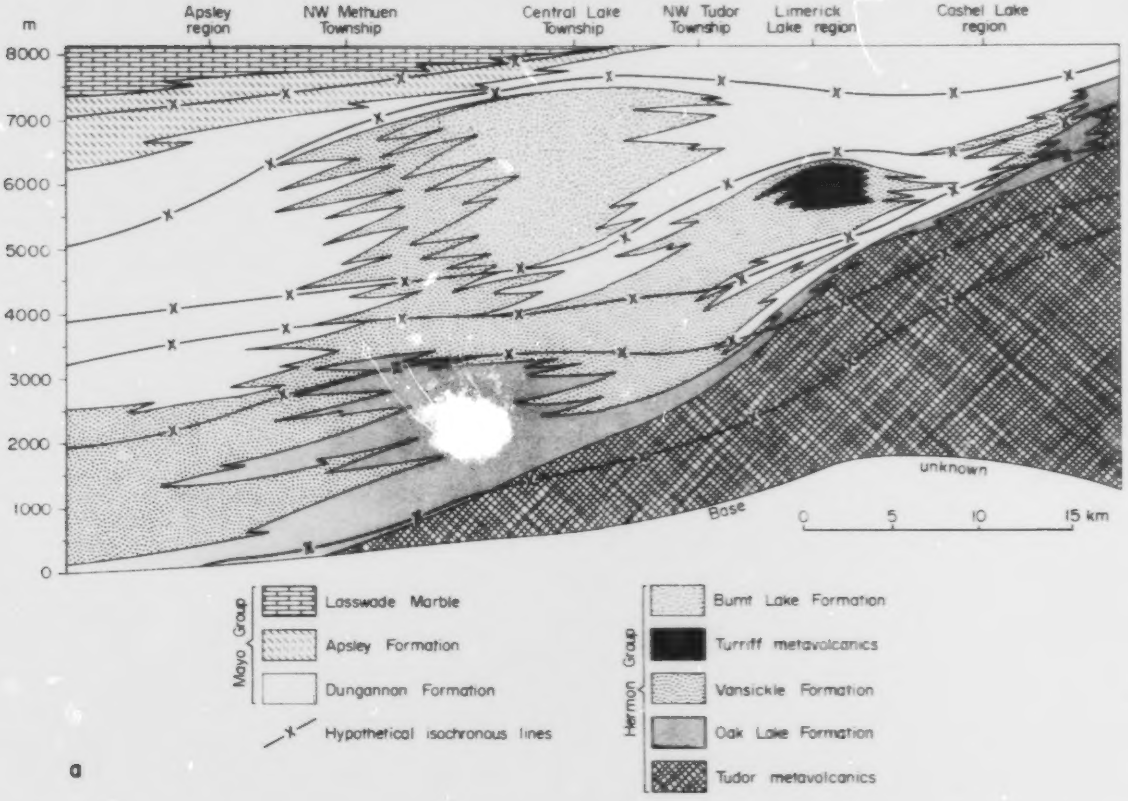

m
Apsley region
NW Methuen Township
Central Lake Township
NW Tudor Township
Limerick Lake region
Cashel Lake region
8000
7000
6000
5000
4000
3000
2000
1000
0
unknown
Base
0
5
10
15 km
Mayo Group
Lasswade Marble
Apsley Formation
Dungannon Formation
Hypothetical isochronous lines
Hermon Group
Burnt Lake Formation
Turriff metavolcanics
Vansickle Formation
Oak Lake Formation
Tudor metavolcanics
a

**Figure 19.56a.** Stratigraphic relationships between formations of the Hermon and Mayo groups in the Belmont Domain (*modified from* Lumbers 1967a, 1967b).

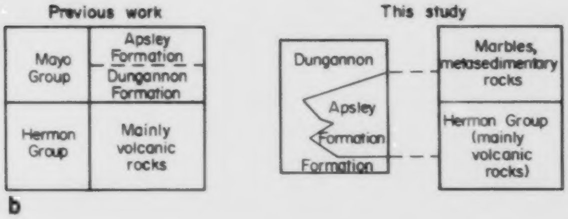

Previous work
This study
Mayo Group
Apsley Formation
Dungannon Formation
Hermon Group
Mainly volcanic rocks
Dungannon
Apsley
Formation
Formation
Marbles, metasedimentary rocks
Hermon Group (mainly volcanic rocks)
b

**Figure 19.56b.** Block diagram showing suggested stratigraphic relations in the western Belmont Domain (*modified from* Easton 1986d, 1987f, 1987g).

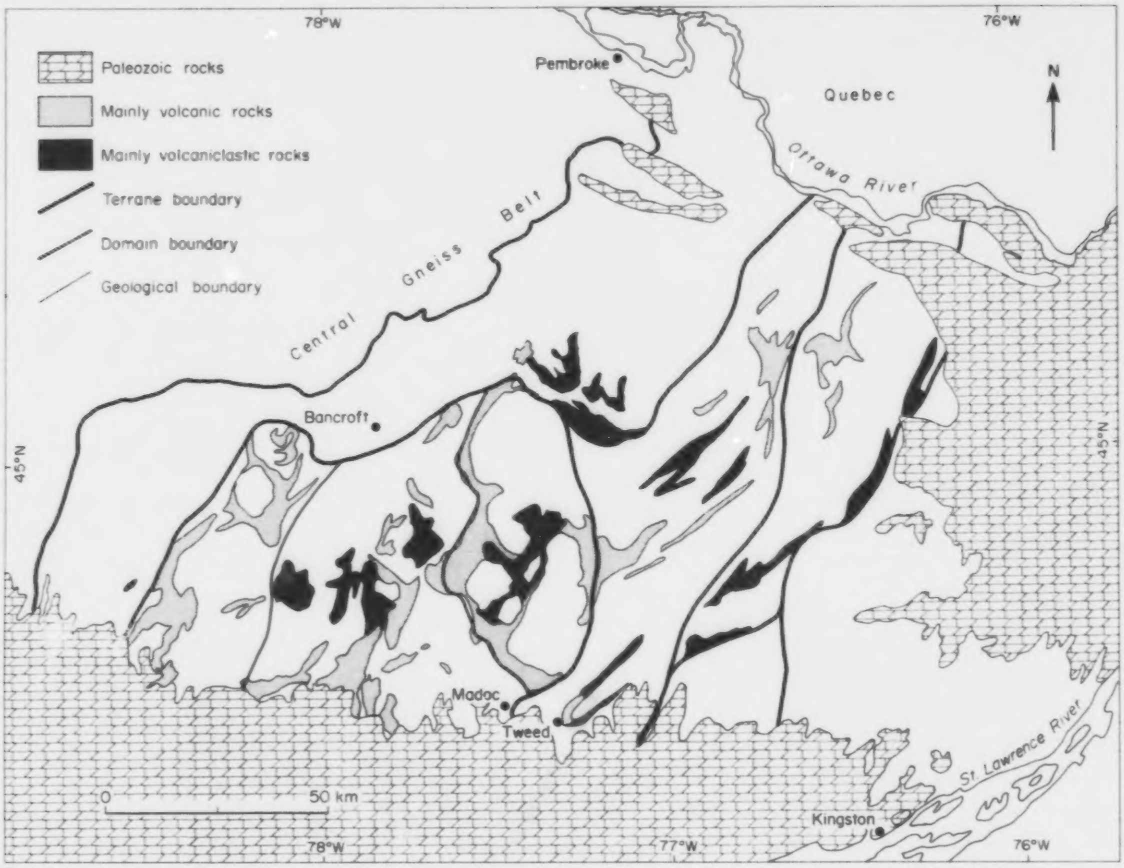
## EARLY TONALITE "DYSART" SUITE (1370 TO 1350 Ma)

This suite is restricted to the northwestern margin of the Bancroft Terrane. Hornblende tonalite, trondhjemite and granodiorite dominate the suite. Metamorphic clinopyroxene is locally present. Dikes of aplite and granite pegmatite are common (Lumbers et al. 1990). Lumbers et al. (1990) report that zircons from the Dysart Tonalite Gneiss fall into the 1370 to 1350 million year range, and the adjacent Redstone Lake Tonalite Gneiss (Hanmer 1988, 1989) is similar in age (U-Pb zircon, 1344$^{+93}_{-32}$ Ma, van Breemen and Hanmer 1986). The geochemistry of the suite is discussed in detail under the section on the Bancroft Terrane.

These rocks may be part of the Central Gneiss Belt and thus, basement to parts of the Haliburton Group. Rocks of *circa* 1350 million year age are present in the Algonquin Terrane, notably the Nobel Granodiorite near Parry Sound, and this age is roughly similar to the age of some rocks within the Parry Sound Terrane (e.g., Whitestone Gabbroic Anorthosite). Rocks of *circa* 1360 to 1340 and 1310 Ma have also been reported from a number of Grenville inliers in the Appalachians (Green Mountains, Vermont, Aleinikoff et al. 1990, Ratcliffe and Aleinikoff 1990, Ratcliffe et al. 1991; Adirondack Highlands, New York, Chiarenzelli and McLelland 1990 and Daly and McLelland 1990). Thus, rocks of the Dysart Suite may be part of a more regional magmatic episode involving dacitic volcanism and tonalitic intrusion.

## NEPHELINE SYENITE SUITE (ca. 1280 TO 1250 Ma)

The nepheline syenite suite includes the nepheline gneisses of the Haliburton–Bancroft region, well known to petrologists. The suite is most extensively developed in the northern part of the Bancroft Terrane (*see* Figure 19.71b), forming a discontinuous zone up to 15 km wide, but



a

**Figure 19.57a.** Distribution of volcanic and volcaniclastic rocks in the Central Metasedimentary Belt.

members of this suite occur in the western Elzevir Terrane (e.g., Blue Mountain Nepheline Syenite); and, as noted in the section on the Central Gneiss Belt, the Bigwood Alkalic Complex and the Calvin Township nepheline syenites may also be related to this suite. In the Bancroft Terrane, several belts of discontinuous bodies are present, with the greatest number and variety being present in the vicinity of Bancroft (*see* Figure 19.71b). Detailed studies on various parts of the suite include Adams and Barlow (1910), von Eckermann (1942), Gummer and Burr (1943, 1946), Satterly (1943, 1945), Thomson (1943), Moyd (1949), Derry (1951), Baragar (1953), Derry and Phipps (1957), Tilley (1958), Gittins (1961, 1967), Hewitt (1961), Tilley and Gittins (1961), Appleyard (1965, 1967, 1969, 1974a), Armstrong and Gittins (1968), Reeve and Anderson (1976), Sylvester and Anderson (1976), Gittins and Miller (1978), Miller (1983, 1984) and Lumbers et al. (1990).

Rocks of the suite are now mainly gneisses that have been considerably deformed and recrystallized. Because of metamorphism and deformation, many early studies concluded that the rocks were metamorphic or metasomatic in origin. However, a few plutons (Trooper Lake–Laronde Creek, Miller 1983; Conroy Marsh, Sebastopol and Highland Lake, Lumbers 1982; Body "B" in Easton 1987a) are relatively unrecrystallized and contain primary igneous textures and mineralogy. Many plutons are elongate, a feature early workers attributed to metasomatism of sedimentary rocks, but which have now been demonstrated to reflect original emplacement as sills, rather than equant bodies (Lumbers et al. 1990).

As noted in greater detail below, the distribution of the roughly 200 million year younger fenite-carbonatite and granite pegmatite suite overlaps that of the nepheline syenite suite. Fenitic alteration has remobilized some of the nepheline syenite suite rocks and served to obscure primary igneous relationships, particularly, in the Bancroft area. Much of the controversy on the origin of the nepheline syenite suite can be attributed to this overlapping of 2 distinct periods of chemically similar magmatism.

Altered zircons in a gneissic nepheline syenite give a minimum age of 1219 Ma (Miller 1983) for this suite, and Lumbers et al. (1990) estimated the age range of the suite as between 1290 and 1250 Ma on the basis of relative age relationships.

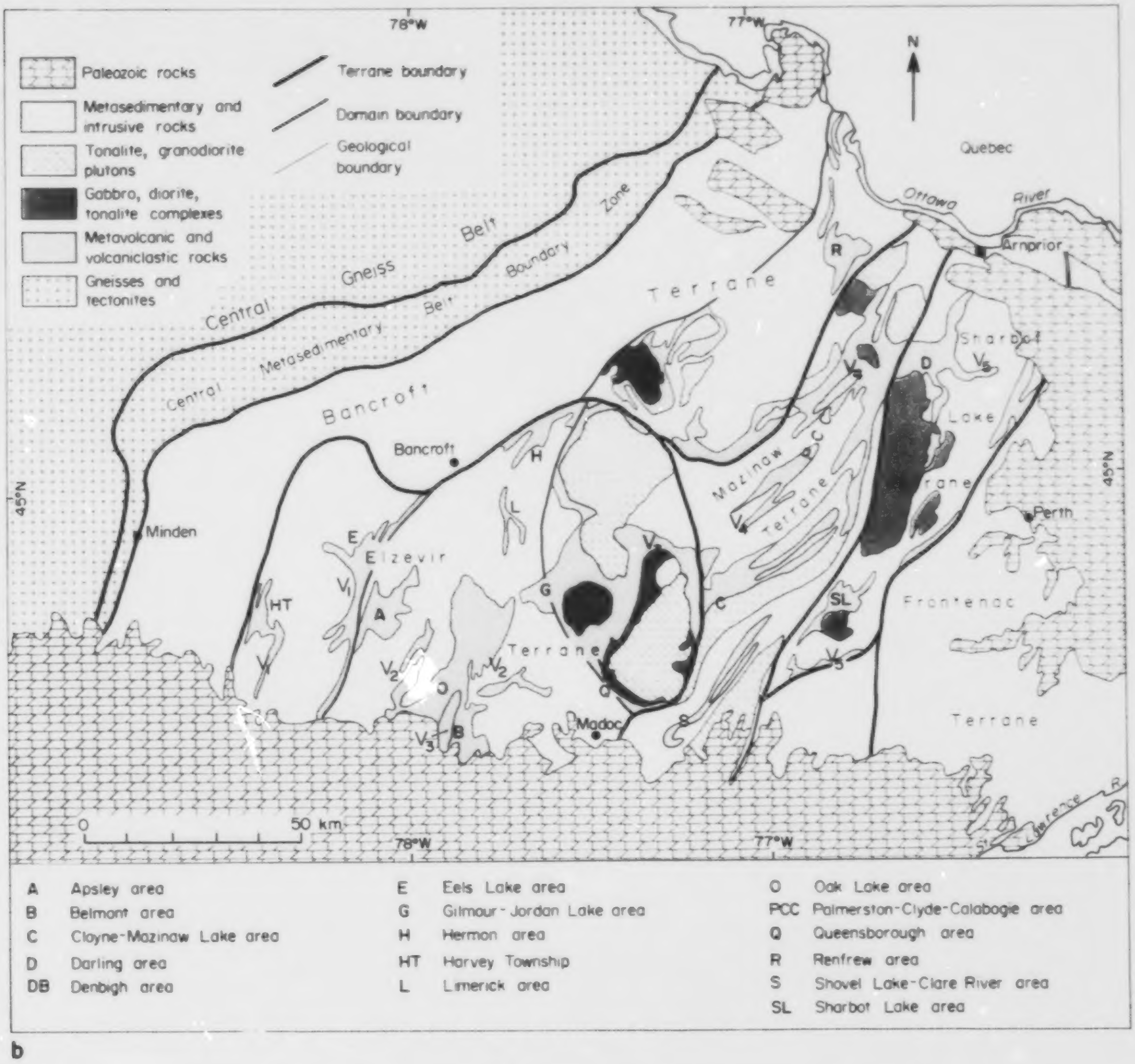

78°W
77°W
N
Paleozoic rocks
Metasedimentary and intrusive rocks
Tonalite, granodiorite plutons
Gabbro, diorite, tonalite complexes
Metavolcanic and volcaniclastic rocks
Gneisses and tectonites
Terrane boundary
Domain boundary
Geological boundary
Quebec
Ottawa River
Arnprior
Central Gneiss Belt
Central Metasedimentary Belt Boundary Zone
Central
Bancroft Terrane
Bancroft
Minden
Elzevir Terrane
Mazinaw Terrane
Sharbot Lake Terrane
Frontenac Terrane
Perth
Madoc
St. Lawrence R.
45°N
0 50 km
V1
V2
V3
V4
V5
A Apsley area
B Belmont area
C Cloyne-Mazinaw Lake area
D Darling area
DB Denbigh area
E Eels Lake area
G Gilmour-Jordan Lake area
H Hermon area
HT Harvey Township
L Limerick area
O Oak Lake area
PCC Palmerston-Clyde-Calabogie area
Q Queensborough area
R Renfrew area
S Shovel Lake-Clare River area
SL Sharbot Lake area
b

**Figure 19.57b.** Distribution of volcanic assemblages within the Central Metasedimentary Belt (*modified from* Easton 1989b). $V_1$ through $V_5$ refer to volcanic-plutonic assemblages described in the text and Table 19.15.

## ANORTHOSITE SUITE (1280 TO 1250 Ma)

Gabbroic and anorthositic rocks of this suite form small plutons spatially associated with rocks of the nepheline syenite suite (e.g., Lumbers 1982; Lumbers et al. 1990), or they occur as thrust sheets within the Denna Lake Structural Complex (e.g., Allsaw Anorthosite, Easton 1987e, 1990b) or the Central Metasedimentary Belt Boundary Zone (e.g., Percy Lake meta-anorthosite, Marmont et al. 1988). A few smaller bodies are found in the Mazinaw, Sharbot Lake and Frontenac terranes. Most of the suite consists of anorthosite and gabbroic anorthosite. Some bodies contain gabbroic and dioritic phases, generally near the margins of the bodies.

Easton (1987e, 1990b) noted that the anorthosite suite rocks in the Bancroft Terrane differ from typical massif-type anorthosites in being characterized by: 1) low abundances of FeO, MgO, $TiO_2$, $P_2O_5$, Zr, Y and V; 2) being locally nepheline normative; and 3) having rare earth element patterns similar to nepheline syenite rocks (Easton 1990b; *see* Figure 19.23). Lumbers (1976b, 1982), Lumbers et al. (1990) and Easton (1987e, 1990b) have suggested a genetic relationship between the anorthosite and nepheline syenite suites in the Bancroft Terrane. Analyses of the anorthosite suite rocks in the Elzevir Terrane are more typical of massif-type anorthosites found within the Grenville Province. Two anorthosite associations are present within the rest of the Central Metasedimentary Belt (Easton 1990b): 1) minor gabbroic anorthosite within differentiated, commonly layered gabbroic intrusions, particularly in the Sharbot Lake Terrane (e.g., Mountain Grove and Lavant

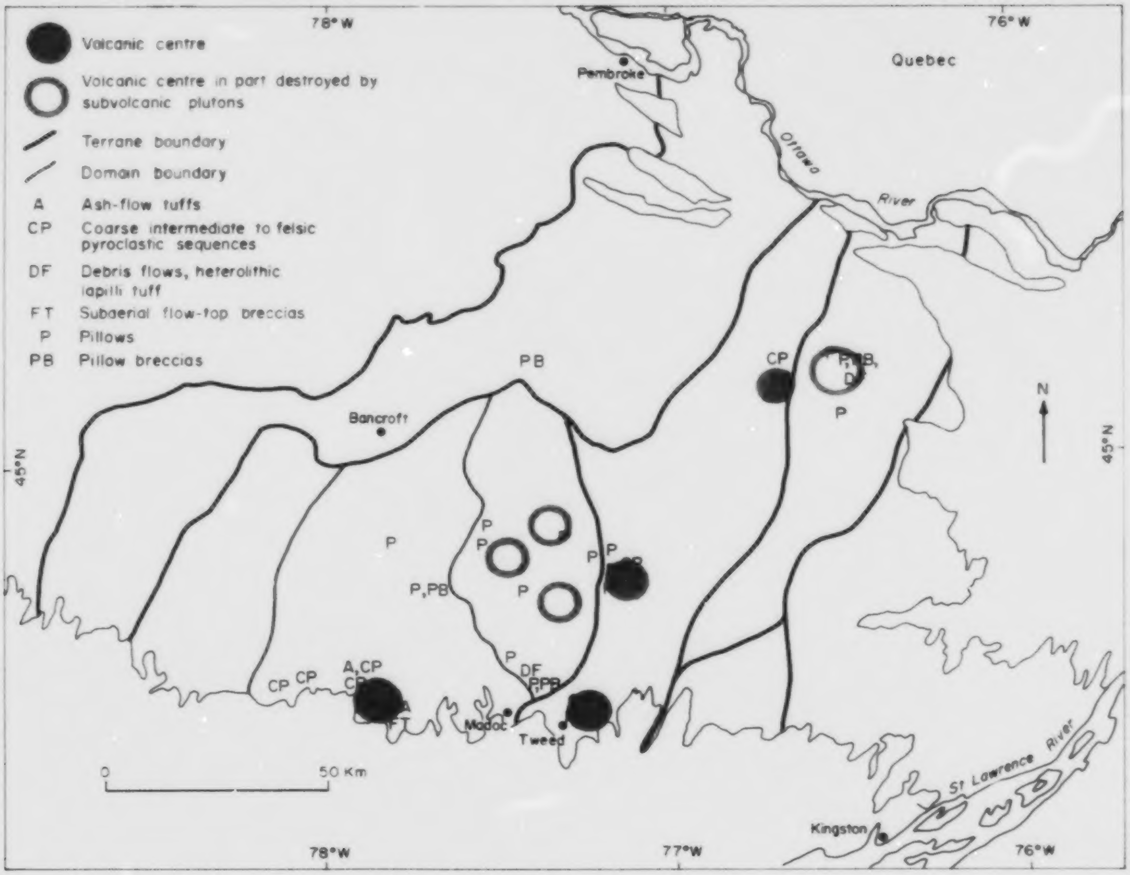

Volcanic centre
Volcanic centre in part destroyed by subvolcanic plutons
Terrane boundary
Domain boundary
A Ash-flow tuffs
CP Coarse intermediate to felsic pyroclastic sequences
DF Debris flows, heterolithic lapilli tuff
FT Subaerial flow-top breccias
P Pillows
PB Pillow breccias
78°W
76°W
77°W
45°N
N
Quebec
Pembroke
Ottawa River
Bancroft
Madoc
Tweed
Kingston
St Lawrence River
0
50 Km
P
PB
CP
A,CP
DF
FT
P,PB

Figure 19.58. Distribution of physical volcanic features in the Central Metasedimentary Belt and location of possible volcanic centres.

gabbros); and 2) anorthosite and gabbroic anorthosite (e.g., Sharbot Lake Terrane). As in the case of the Lavant Suite (*see* Diorite-Gabbro "Lavant" Suite), rocks of this suite are not well-dated. Members in the Bancroft Terrane are *circa* 1290 to 1250 million years in age if they are related to the nepheline syenite suite.

## OLDER "KILLER CREEK" GABBRO SUITE (LESS THAN 1275 Ma)

Large gabbro intrusions containing associated hornblendite and anorthositic gabbros, with primary igneous layering, constitute the older gabbro suite. These intrusions are best preserved in the Grimsthorpe Domain (Easton and Ford 1990a) but also occur in the western Mazinaw Terrane (Easton and Ford 1991).

The largest body of this suite is the Killer Creek Gabbro (Easton and Ford 1990a, in prep.) located in the eastern Grimsthorpe Domain. Ilmenite, chromite and magnetite are locally concentrated in hornblendite pods within the intrusion. The Killer Creek Gabbro is intruded by *circa* 1270 million-year-old Elzevir Tonalite. Preliminary geochemical study of the Killer Creek Gabbro (Easton, unpublished data) indicates elevated concentrations of $TiO_2$ (greater than 1.7%), $P_2O_5$ (greater than 0.6%) and Cu (greater than 450 ppm), and may be indicative of generation in an island-arc setting. This is in marked contrast to the anorthosite suite, and the 2 are not likely comagmatic. Preliminary data suggests that copper-nickel magmatic sulphide deposits appear to be more abundant in members of this suite than in the younger Lavant Suite.

## TONALITE "ELZEVIR" SUITE (ca. 1280 TO 1270 Ma)

This suite is restricted to the Bancroft and Elzevir terranes and is most extensive within the Elzevir Terrane, where it is closely associated with the main belts of volcanic rocks. Volumetrically, it is the most abundant plutonic rock suite within the Central Metasedimentary Belt (Lumbers et al. 1990).

The suite consists of several rock types, including trondhjemite, tonalite and granodiorite. Lumbers et al. (1990) suggest that trondhjemite constitutes the bulk of the suite, however, using Streckeisen's (1976) nomenclature, tonalites are the dominant rock type, and this is the usage in this paper. Albite granite and alkali feldspar granite occur as late dikes cutting plutons of the suite and as fine-grained satellite plutons associated with calc-alkalic volcanic rocks (Lumbers et al. 1990).

The geochemistry of several bodies of this suite have been studied in detail (Pride and Moore 1983; Somers 1984; Wu 1984; Easton 1987a, 1987e; Wu et al. 1991). Most of the tonalite and trondhjemites are of the high-alumina type, and are biotite rich, with hornblende being rare or absent. Chemically, they fall between calc-alkalic and trondhjemite trends. Lumbers et al. (1990) report zircon ages of 1280 to 1270 Ma for rocks of this suite, overlapping or slightly younger than the tholeiitic volcanic rocks which they commonly intrude. An origin by partial melting of tholeiitic basalts in either an island-arc or mantle-activated rift is suggested by Lumbers et al. (1990) and is consistent with low initial strontium ratios (Bell and Blenkinsop 1980) and lack of zircon inheritance in these rocks.

## DIORITE-GABBRO "LAVANT" SUITE (1250 TO 1230 Ma)

Lumbers et al. (1990) have described a diorite-gabbro suite throughout the Elzevir and Frontenac terranes that is spatially associated with the volcanic and late tonalite suites of the Central Metasedimentary Belt. Recent geochronology suggests that the Chenaux, Raglan–Boulter–Mallard Lake, and Trooper Lake gabbroic intrusions within the Bancroft Terrane are part of this suite (McEachern et al. 1992), although lithologically, the Bancroft Terrane gabbroic intrusions are most similar to the Killer Creek Suite. According to Lumbers et al. (1990), plutons of the Lavant Suite are composed mainly of diorite or gabbro or both. Locally, pyroxenite or gabbroic anorthosite phases are present, and late dikes of tonalite, granodiorite, and more rarely syenite, occur in the marginal zones of the plutons.

## ALASKITE "METHUEN" SUITE (1250 TO 1240 Ma)

Alaskitic granites of this suite are the most widely distributed, and the second most voluminous plutonic suite (Lumbers et al. 1990) within the Central Metasedimentary Belt, occurring throughout the Bancroft and Elzevir terranes. Alaskitic granites of similar age are present in the Adirondack Lowlands portion of the Frontenac Terrane (deLorraine and Carl 1986; Chiarenzelli and McLelland 1990; McLelland et al. 1992).

The suite consists of leucocratic, pink monzogranites containing biotite and ferrohastingsite with accessory titanite, fluorite and magnetite. The granites are mainly metaluminous to marginally peraluminous, but peralkaline varieties do occur (e.g., Deloro and Cheddar granites; Lumbers et al. 1990). Lumbers et al. (1990) note that most plutons are subalkalic, and contain between 70 and 78% $SiO_2$, but that rocks of the suite containing less than 70% $SiO_2$ are mainly alkalic. Lumbers et al. (1990) classify them as anorogenic granites (*after* Anderson 1983), in that they contain relatively high FeO/FeO+MgO ratios, less $Na_2O$ and $SiO_2$ and more CaO then calc-alkalic granites. Lumbers et al. (1990) note that this magmatic episode marks a major departure from the earlier tholeiitic and calc-alkalic magmatism in the Central Metasedimentary Belt.

The 6 dated plutons of the suite lie in the range 1250 to 1240 Ma, which overlaps the time span of the diorite-gabbro plutonic and calc-alkalic volcanic suites. Rocks of the alaskite suite contain low initial strontium ratios (less than 0.7040, Heaman et al. 1986), and Lumbers et al. (1990) suggest a derivation by small degrees of melting of slightly older calc-alkalic granitoids (e.g., Elzevir Suite).

## MONZONITE-SYENITE-GRANITE "GANANOQUE" SUITE (1180 TO 1165 Ma)

This suite is only present within the Frontenac Terrane (Wynne-Edwards 1967a; Currie and Ermanovics 1971; Shieh 1985; Lumbers et al. 1990; Marcantonio et al. 1990) and consists mainly of massive syenite and granitic rocks largely unaffected by deformation and recrystallization. The plutons cut mylonitic fabric in their host rocks but locally contain post-Grenvillian mylonitic zones (e.g., Perth Road mylonite zone, Barclay 1985, Carmichael et al. 1987). Rocks of the suite are rich in mesoperthite, sodic oligoclase and biotite, and many plutons contain minor clinopyroxene. Monzonite and syenite plutons are equally abundant. Syenogranite and monzogranite are minor phases in most plutons, with the exception of the Lyndhurst and Battersea monzogranites. Gabbro and diorite xenoliths are common as xenoliths in the felsic plutons of this suite.

Wynne-Edwards (1967a) and Sauerbrei (1966) divided rocks of the suite into 2 types: 1) the Frontenac-type, forming discrete plutons, ranging in composition from syenite to granite; and 2) the Rockport-type, consisting mainly of biotite granite and monzonite intimately associated with migmatites. Plutons of both types intrude and include one another, although Wynne-Edwards (1967b) notes that the Rockport Granite is generally conformable with the country rocks. The major Rockport- type pluton is the $1173^{+4}_{-3}$ million-year-old Rockport Granite (van Breemen and Davidson 1988b). Frontenac-type plutons are slightly younger, with ages ranging from 1162 to 1176 Ma (van Breemen and Davidson 1988b; Marcantonio et al. 1990). Nd-model ages of the plutons studied by Marcantonio et al. (1990) range from 1.48 to 1.34 Ga, notably older than their intrusion ages, suggesting crustal contamination of the magmas.

Marcantonio et al. (1990) studied the chemistry of the suite and classified these rocks as I-type plutons with a $Al_2O_3/(Na_2O+K_2O+CaO)$ mole ratio of less than 1 for all plutons and $Na_2O$ contents of greater than 3.2 wt %. All are diopside normative except for the Crow Lake Monzonite. Marcantonio et al. (1990) favoured a petrogenesis involving mixing of a basic crustal source, with contamination by surrounding metasedimentary rocks having a Nd model age of 1790 Ma and interaction with marble-derived fluids.

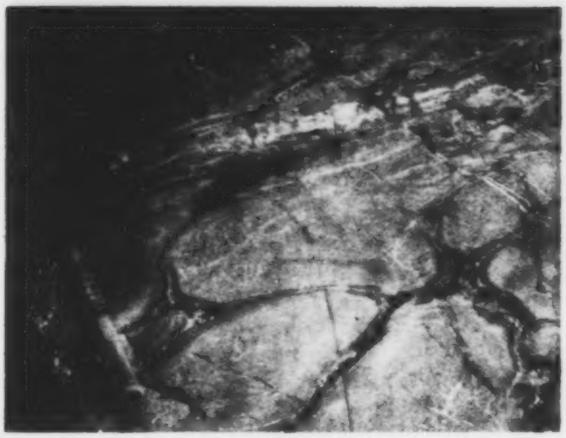


**Figure 19.59a.** Pillowed mafic volcanic rocks, junction of highways 41 and 506, near Cloyne. Mazinaw Terrane. Hammer handle is 30 cm long.

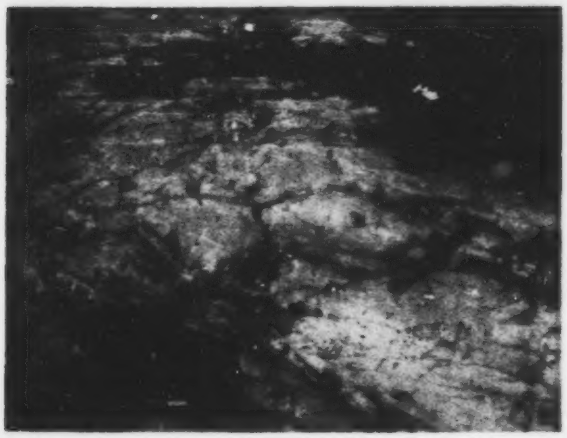


**Figure 19.59b.** Pillowed mafic volcanic rocks, Grimsthorpe Domain. Note thin selvages and lack of hyaloclastite between pillows. Hammer handle is 30 cm long.

**Figure 19.59c.** Brecciated mafic volcanic rock of the Grimsthorpe Domain. Origin of this texture is unknown, but texture is common throughout the domain. Hammer handle is 30 cm long.

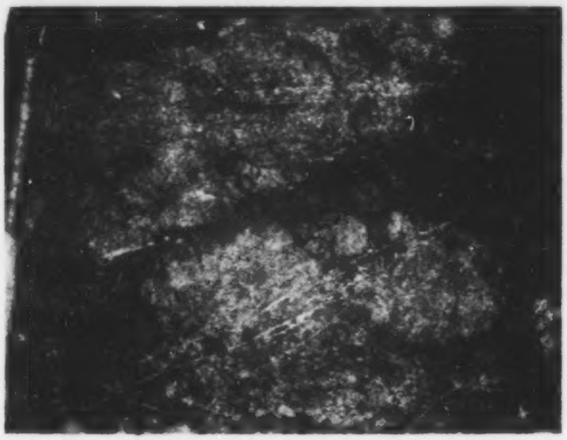


**Figure 19.59d.** Brecciated mafic volcanic rock of the Grimsthorpe Domain. Origin of this texture is unknown, but texture is common throughout the domain. Hammer handle is 30 cm long.

## SYENITE "SKOOTAMATTA" SUITE (1090 TO 1075 Ma)

This plutonic suite is present in the Bancroft, Elzevir and Frontenac terranes, however, it has not been reported in the Frontenac Terrane south of the Rideau–Canoe Lake fault. This suite has been studied by numerous workers, including Wynne-Edwards (1967a, 1967b), Corriveau (1989, 1990), Corriveau et al. (1990), Shieh (1985), Marcantonio et al. (1990) and Lumbers et al. (1990).

Marcantonio et al. (1990) classified them as I-type plutons based on geochemical evidence. The mole ration $Al_2O_3/(Na_2O+K_2O+CaO)$ is less than 1 for all plutons, $Na_2O$ contents are greater than 3.2 wt % and all are diopside normative. Skootamatta Suite plutons have less $SiO_2$ and CaO, but more $Al_2O_3$, $K_2O$ and $Na_2O$ than the monzonite-syenite-granite suite plutons (Lumbers et al. 1990; Marcantonio et al. 1990). Marcantonio et al. (1990) favoured a petrogenesis involving mixing of a basic crustal source, with contamination by surrounding metasedimentary rocks having a Nd model age of 1790 Ma, and interaction with marble-derived fluids, similar to the origin of the monzonite-syenite-granite suite plutons.

Corriveau (1989, 1990) studied plutons of this suite in western Quebec which are more differentiated than equivalent plutons in Ontario. She classified them as a potassium-rich alkalic and shoshonitic suite, characteristic of Group III ultrapotassic rocks (Foley et al. 1987), which form in island-arc environments.

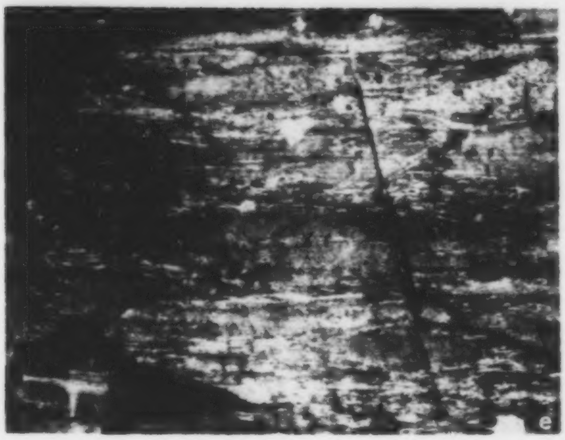


**Figure 19.59e.** Tuff breccia consisting of mafic to intermediate composition fragments in an ash matrix, possibly a debris flow. Sharbot Lake Terrane. Hammer head is 17 cm across.

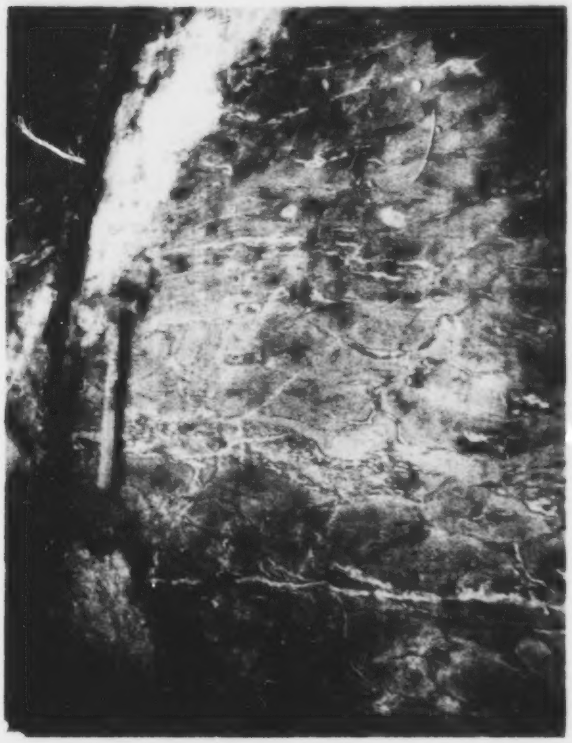


**Figure 19.59f.** Pillow breccia composed of small pillow fragments and hyaloclastite, Raglan area, Bancroft Terrane. Hammer handle is 30 cm long.

## CARBONATITE AND FENITE SUITE (1070 TO 1040 Ma) AND GRANITE PEGMATITE SUITE (1050 TO 1030 Ma)

Lumbers et al. (1990) have documented a suite of carbonatitic rocks and associated fenites ranging in age from 1070 to 1040 Ma (Lumbers et al. 1990). The fenites are characterized by patches and veinlets of alkali pyroxene and amphibole and occur in a band running the length of the Bancroft Terrane paralleling the CMBBZ. The fenitizing fluids introduced sodium, potassium, calcium, iron, titanium, phosphorus, fluorine, chlorine, uranium, thorium, rare earth elements, zirconium, barium and molybdenum into the country rocks (Lumbers et al. 1990). The carbonatite suite includes the so-called calcite "vein dikes" of the Bancroft region (*see* references *under* Bancroft Terrane). Associated with these rocks are a variety of granitoid pegmatites, some of which are uranium-, molybdenum- and rare element-bearing. Although late pegmatites occur throughout the Central Metasedimentary Belt, they are most abundant in the general vicinity of the fenite-carbonatite belt, particularly the uranium, molybdenum and rare-element pegmatites. Lentz (1991a) provides many descriptions of the fenites and their associated pegmatites and carbonatites. Several hypotheses have been proposed for the origin these pegmatites, skarns and veins, as summarized by Lentz (1991a).

Lentz (1991a) contends that the spatial, temporal and geochemical relationships favour a magmatic-hydrothermal origin for the skarns and veins related to the emplacement of granitic pegmatites at intermediate crustal levels. The magmatic-hydrothermal fluids were derived from the crystallizing granitic pegmatites at volatile-phase saturation and were responsible for skarn formation. The chemically modified hydrothermal fluid produces "fenite-like" alteration due to calcium, magnesium, alkali and iron metasomatism during deposition of calcite, apatite and fluorite in veins formed along older structures. Lentz (1991a) suggested that this occurred in the abyssal environment (at roughly 500°C and 4 kb) based on thermochronologic studies (Berger and York 1981a, 1981b).

Goad (1990) studied the pegmatites of the Central Metasedimentary Belt to evaluate their rare-element mineral potential and to classify them into distinct pegmatite fields. He concluded that the pegmatites could not be classified into distinct geochemical groups, and that all the pegmatites studied fell into the Mica-Ceramic type pegmatites of Gordiyenko (1976) and consequently had limited rare metal-bearing potential. Figures 19.71e and 19.20d show the distribution of the main pegmatite fields in the Grenville Province in Ontario, based largely on the classifications of Hewitt (1967b) and Goad (1990). Goad (1990) attributed pegmatite genesis to partial melting related to upper amphibolite- to granulite-facies regional metamorphism.

Lumbers et al. (1990) relate all pegmatites in the Central Metasedimentary Belt as well as the zone of carbonatite and fenites in the Bancroft Terrane to a regional metasomatic event generated by high-temperature alkaline fluids which rose through the crust along fractures.

The carbonatites are generally small, elongate bodies, less than 1 m wide, but a few, such as the Silver Crater carbonatite, outcrop as circular masses, generally less than 100 m across. Zircon ages from the carbonatites and fenites from the zone reported in Lumbers et al. (1990) lie in the range 1070 to 1038 Ma, and ages of the fenites, including

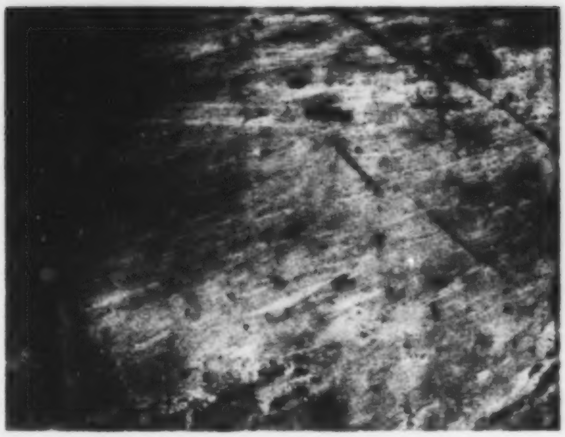


**Figure 19.60a.** Lapilli fragments in felsic metavolcanic rocks of the Mazinaw Terrane north of Clyde Forks. Hammer handle is 30 cm long.

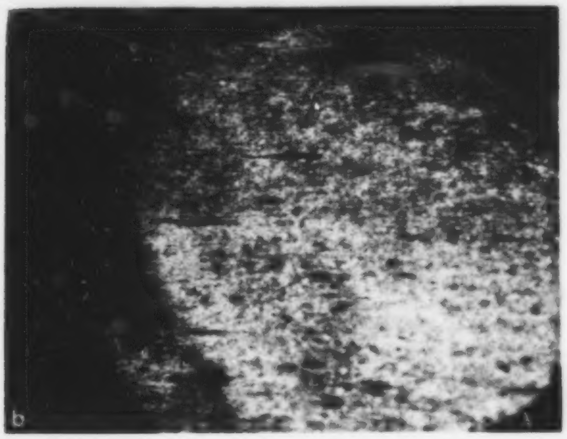


**Figure 19.60b.** Possible ash-flow tuff, felsic metavolcanic rocks of the Belmont Terrane. Lens cap is 5 cm across.

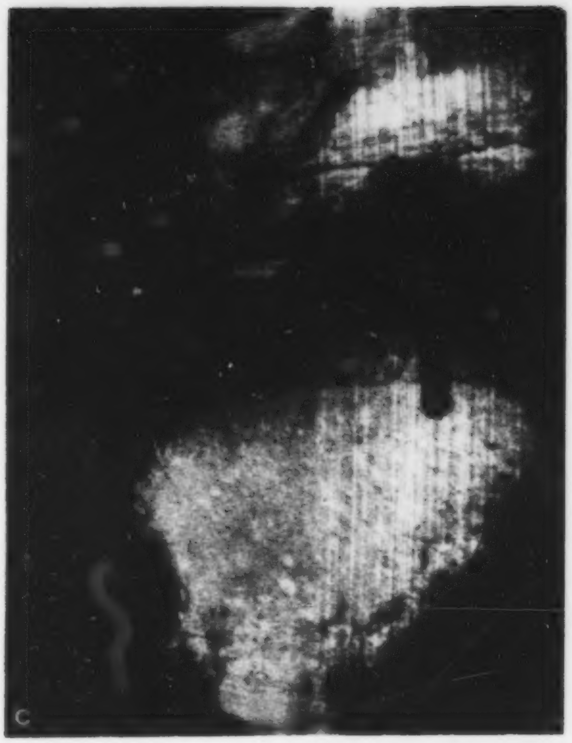


**Figure 19.60c.** Doubly graded bed within the Apsley Formation, Belmont Domain. Bed at base fines upward, and sequence of beds also fines upward. This unit may be a subaqueous pyroclastic flow (c.f., Fiske and Matsuda 1964). Hammer handle is 30 cm long.

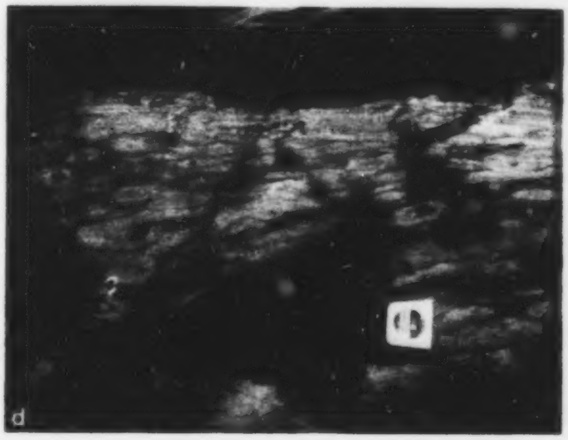


**Figure 19.60d.** Monolithic, tuff-breccia bed within the Apsley Formation, Belmont Domain. Tape measure is 6.5 cm long.

metamorphic zircons from fenitized nepheline syenite suite rocks studied by Miller (1983, 1984), span a similar range from 1072 to 1044 Ma.

The models proposed by Lentz (1991a) and Lumbers et al. (1990) rely on magmatic-hydrothermal fluids related to the emplacement of granite pegmatite veins to form the skarn assemblages and to introduce the mineralizing fluids. The main difference is whether the alteration is termed "fenite" and if the calcite veins are "carbonatites" or simply the products of hydrothermal activity.

In either case, this "carbonatite" intrusive event occurred after "Grenvillian" metamorphism and is not related to the Ottawan Orogeny. Timing of the event may coincide with the development of extensional shear systems in the region (dated at ca. 1040 Ma in the Central Gneiss Belt, Nadeau and van Breemen 1990). The localization of the most intense zone of pegmatite injection, fenitization and carbonatite emplacement near the Bancroft Terrane–Central Metasedimentary Belt Boundary Zone may be structurally controlled.

## Mineral Deposits

Carter (1984), Carter and Colvine (1985), Sangster (1970) and Sangster and Bourne (1982) have provided

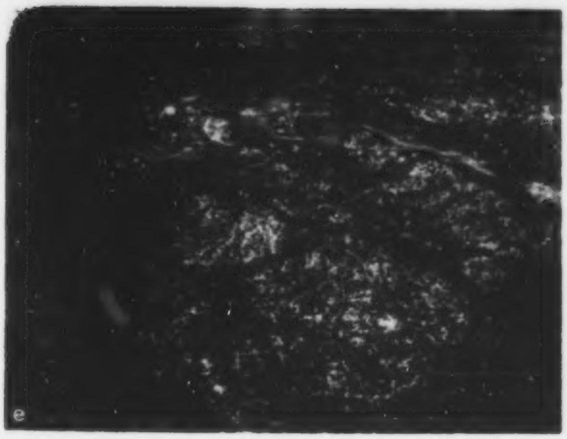


**Figure 19.60e.** Matrix-supported, volcanic conglomerate of the Grimsthorpe Domain containing large gabbro clasts. Clasts are derived from older gabbro unit shown in Figure 19.81 which represents one of the oldest rocks in the Central Metasedimentary Belt. Hammer handle is 30 cm long.

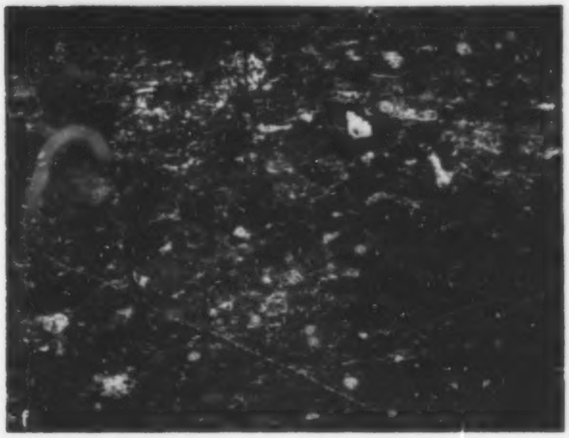


**Figure 19.60f.** Clast-supported, volcanic conglomerate of the Grimsthorpe Domain containing a variety of volcanic and volcanic sediment clasts. Field of view is 20 cm across.

descriptive classifications of metallic mineral deposits in the Central Metasedimentary Belt. Carter (1984) and Carter and Colvine (1985), in particular, noted the importance of stratigraphy and rock type in controlling the distribution of metallic mineral deposits in the Central Metasedimentary Belt, including some types of magmatic and contact metamorphic deposits. Easton et al. (1986) extended this classification to include industrial mineral deposits.

The classification of Carter (1984) served as the groundwork for the "Metallogeny of the Grenville Province" chapter (this volume). The concept of stratigraphic and lithologic control on mineralization is emphasized herein, as the major differences in stratigraphy, plutonism and tectonism between the Central Metasedimentary Belt terranes and domains are reflected in the distribution of mineral deposits. Some types of mineral deposits are absent from some domains because of the absence of critical lithologies or depositional and tectonic events which played an important role in the mineralization process. This does not mean that all deposits are stratiform or strata-bound, rather only that the specific circumstances controlling mineralization are strongly influenced by host rocks. For example, the unconformity between the Flinton Group and the Grenville Supergroup (specifically the Mazinaw Group) has served as a locus for mineralization. It is still not clear what role the Flinton Group plays in the mineralization (e.g., good cap rock, source of metals, compositional control on mineralizing fluids, structural control related to the depositional environment of the Flinton, supergene enrichment of underlying strata during weathering or fluid movement during subsequent metamorphism, or some combination of the above), however, the contact between the Flinton Group and the Grenville Supergroup can be targeted for exploration for specific mineral deposits which have no direct counterpart elsewhere in the Central Metasedimentary Belt. Thus, an understanding of the lithologic diversity and tectonic history of the various terranes and domains is relevant to mineral exploration in the Central Metasedimentary Belt. The distribution of types of mineral deposits between the different terranes and domains is summarized in diagrammatic form in the "Metallogeny of the Grenville Province" chapter (this volume).

Many workers treat industrial mineral deposits separately from metallic mineral deposits. In the Central Metasedimentary Belt, however, stratigraphy plays a major role in localizing these deposits as well. For example, the Flinton Group is the major target unit for deposits of sillimanite, kyanite, muscovite and garnet, mainly as it is the only aluminous-rich metasedimentary unit within the Central Metasedimentary Belt. Finding this unit is, thus, essential in locating an industrial mineral prospect for these minerals, with secondary factors being purity, appropriate grain size and logistical considerations. These topics are further discussed in "Metallogeny of the Grenville Province."

Building stone (ashlar, facing, flagstone, ornamental stone) also shows similar stratigraphic controls, but secondary factors such as fracture pattern, proximity to transportation networks, ease of quarrying and so on are probably more critical factors in determining the marketability of these products. To the average geologist, many pluton types in the Central Metasedimentary Belt might seem to be suitable and attractive building stones; however, when all these factors are weighed, few building stone prospects in the Central Metasedimentary Belt can be identified (*see* LeBaron et al. 1990). Of all the various building stone uses, flagstone prospects are probably the most easily identified on a regional scale, as they are restricted to ductile deformation zones subjected to middle to upper amphibolite-facies metamorphism. Thus, domain and terrane boundaries in the Central Gneiss Belt and the CMBBZ are the primary target areas for flagstone deposits.

Numerous inventories of industrial and metallic mineral deposits in the Central Metasedimentary Belt are available as detailed in the "Metallogeny of the Grenville

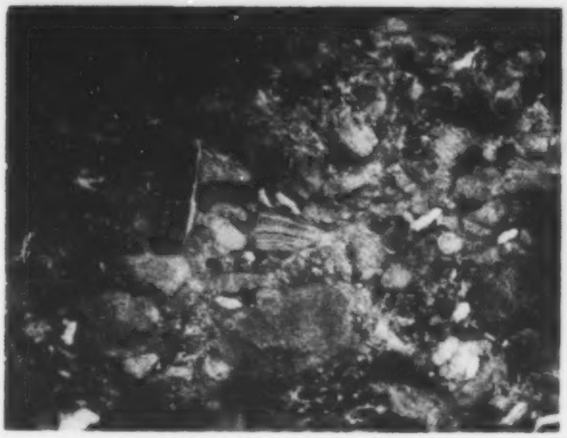


**Figure 19.61a.** Matrix- to clast-supported debris flow containing a variety of volcanic clasts from Belmont Lake, Belmont Domain. Knife is 9 cm long.

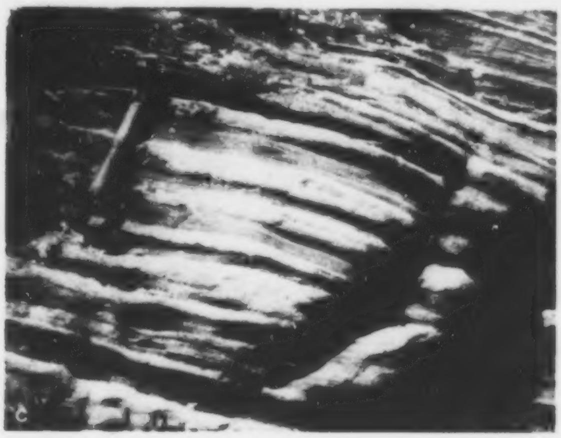


**Figure 19.61c.** Dacitic ash and calc-silicate couplets within the Apsley Formation, Highway 28, north of Apsley. Couplets are probably deposited by turbidity currents, with upper parts also receiving carbonate detritus. Hammer handle is 30 cm long.

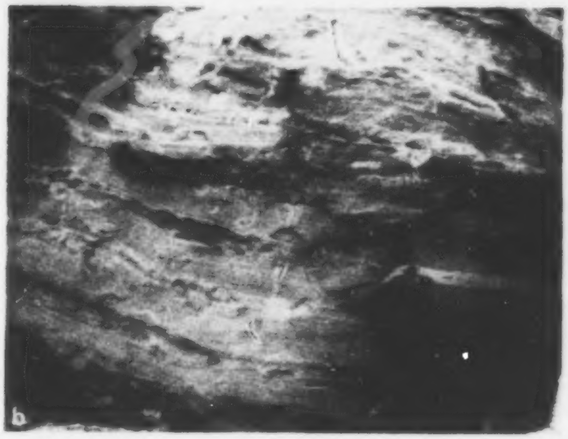


**Figure 19.61b.** Turbidite beds within the distal facies of the Apsley Formation, Chandos Lake. Lens cap is 5 cm across.

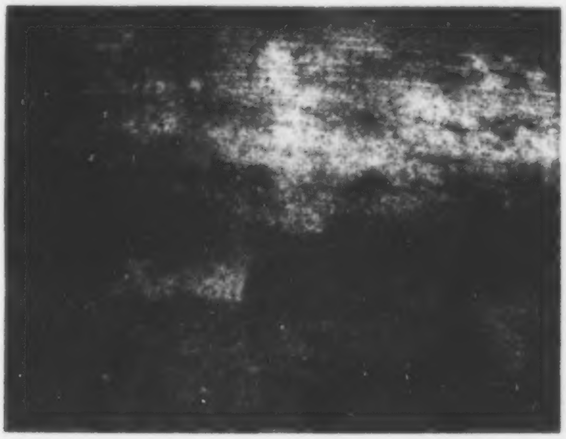


**Figure 19.61d.** Scouring at base of ash bed within the Apsley Formation, Highway 28, north of Apsley. Pen is 13.5 cm long.

Province" (this volume). In addition, some of the more significant deposits of both are described briefly throughout the remainder of this chapter under the descriptions of the different domains and terranes.

# Terrane Descriptions

## BANCROFT TERRANE

### Introduction

Most of the southern part of the Bancroft Terrane has been mapped recently at 1:50 000 scale or better (Bright 1987b; Easton 1987a, 1987d, 1987e, 1988g, 1990a). The mapping done in the rest of the terrane is considerably older (Adams and Barlow 1910; Quinn 1952; Baragar 1953; Hewitt 1954, 1955, 1959, 1961; Hewitt and James 1956; Quinn et al. 1956; Evans 1964; Appleyard 1967, 1974b; Themistocleous 1981a, 1981b; Lumbers 1982; Masson 1982a, 1982b; Breaks and Thivierge 1985). As a consequence of this distribution of mapping, this chapter will focus on the better known southern part of the Bancroft Terrane.

The boundaries of the Bancroft Terrane, particularly with the Elzevir Terrane to the southeast, are not well-defined. The boundary with the Central Metasedimentary Belt Boundary Zone is probably best taken as the northern and western limit of major marble belts. The boundary used herein is illustrated in Figure 19.51 and, in the southwest and Bancroft area, is essentially as in Davidson (1986a) and Moore (1982). As is the case for the Sharbot Lake area, the Bancroft Terrane could be regarded simply as a domain of the Elzevir Terrane.

True metavolcanic rocks are rare to absent in the Bancroft Terrane, which is dominated by calcitic

**Table 19.15.** Characteristic features of volcanic and volcanic-plutonic assemblages in the Central Metasedimentary Belt.

| Code | Major Rock Types | Age | Tectonic Setting |
|---|---|---|---|
| $V_1$ | Basalt, dacite-rhyolite, pyroclastic rocks, related volcaniclastic rocks, tholeiitic to slightly alkalic chemistry.<br>Mineralization: pyrite; rusty schists and black shales associated with the metavolcanic rocks locally show a variety of metal enrichments, including gold, copper, zinc.<br>May be distal equivalent of assemblage $V_2$. | ?? | ?? |
| $V_2$ | Basalt, andesite, dacite, rhyolite, abundant felsic pyroclastic rocks, related andesitic and dacitic volcaniclastic rocks (e.g., Apsley Fm.), tholeiitic (mafic rocks) and calc-alkalic (intermediate to felsic rocks) chemistry. Associated gabbro and granite plutons. Includes classic Oak Lake and Burnt Lake fms.<br>Mineralization: pyrite, copper-zinc; rusty schists and black shales associated with the metavolcanic rocks locally show a variety of metal enrichments, including gold, copper, zinc, lead, silver. | volcanism<br>1260–1248 Ma<br>plutonism<br>1245–1240 Ma | back-arc and/or arc |
| $V_3$ | Basalt, gabbro sills, gabbro and pyroxenite plutons, mafic volcaniclastic sedimentary rocks, tholeiitic chemistry.<br>Associated with gabbro-diorite-tonalite complexes (e.g., Lingham Lake Gabbro-Diorite Pluton) and tonalite-granodiorite plutons (e.g., Elzevir); pillowed flows typically have very thin selvages. Includes much of the classic Tudor Fm.<br>Mineralization: gold, talc, copper, potential for copper-nickel, actinolite. | volcanism<br>1290–1285 Ma<br>plutonism<br>1280–1270 Ma | basal arc? and/or oceanic? |
| $V_4$ | Basalt, andesite, dacite, rhyolite, abundant pyroclastic rocks, volcaniclastic sedimentary rocks, and quartzofeldspathic metasedimentary rocks; tholeiitic basalts and calc-alkalic, intermediate to felsic rocks.<br>Contact relations with assemblage $V_3$ unknown—may overlie it stratigraphically.<br>Mineralization: copper-zinc-pyrite. | <1270 Ma? | arc and/or back arc? |
| $V_5$ | Basalt, gabbro sills, mainly flows, minor mafic pyroclastic rocks, some exhalative rocks including black shales, sulphide-facies iron formation; pillowed flows typically have very thin selvages.<br>Associated with gabbro-diorite intrusions; could be similar to $V_3$ but is not associated with large tonalite-granodiorite plutons.<br>Mineralization: gold (particularly in deformed rocks), copper-zinc pyrite; many zinc deposits are associated with dolomite marbles overlying the metavolcanic rocks. | ?? | ?? |

and dolomitic marbles, commonly graphitic, and quartzofeldspathic gneisses derived from siliceous, clastic sedimentary rocks. In the east, "para-amphibolites", probably representing volcaniclastic sediments are abundant (e.g., Evans 1964), and similar volcaniclastic metasedimentary rocks and distally derived metavolcanic rocks increase in abundance in the southwestern part of the terrane (Easton 1987a). Plutonic rocks consist of tonalite-granodiorite and gabbro-diorite-syenite-granite complexes. Syenitic and nepheline syenitic gneisses are widespread, as are metasomatic rocks (skarns and pegmatites) related to the carbonatite-fenite, magmatic- metasomatic event. The Bancroft Terrane is structurally complex, dominated by marble-breccia zones and abundant high-strain zones, and has been subjected to upper amphibolite-facies metamorphism.

## Bancroft–Elzevir Terrane Boundary

In contrast to other Central Metasedimentary Belt terranes, the placement of the Bancroft–Elzevir terrane boundary is unclear. In part, this is because in defining the Bancroft Terrane, Moore (1982) stated that the Bancroft and Elzevir terranes interfingered, whereas all other terrane boundaries were abrupt and sharp. Other workers have outlined the boundary using major structural features, such as part of the McArthurs Mills and the York River shear zones (e.g., Bartlett et al. 1984; Davidson 1986a).

Moore (1982) defined the Bancroft Terrane as an essentially carbonate basin, with a distinctive plutonic suite of undersaturated rocks. This definition places the boundary of the terrane west of the Harvey-Cardiff Arch, north of the York River shear zone and north of a northeast-trending line stretching from Quadeville to Haley Station. Using this definition, much of the area north of the Abinger Granite (north of the Mazinaw Terrane in Figure 19.51a) and northeast of the Weslemkoon Tonalite can be considered part of the Bancroft Terrane, as suggested by McEachern et al. (1990, 1992), McEachern (1990) and Easton and Ford (1991). This boundary is roughly 30 to 50 km farther south of previous placements of the boundary (cf., Moore 1982; Davidson 1986a).

In the south and central part of the Bancroft Terrane and north of the Abinger Granite, Moore's (1982) lithologic boundary also coincides roughly with the southern limit of a zone characterized by a prominent southeast-plunging lineation (Hewitt 1962; Hanmer and Ciesielski 1984;

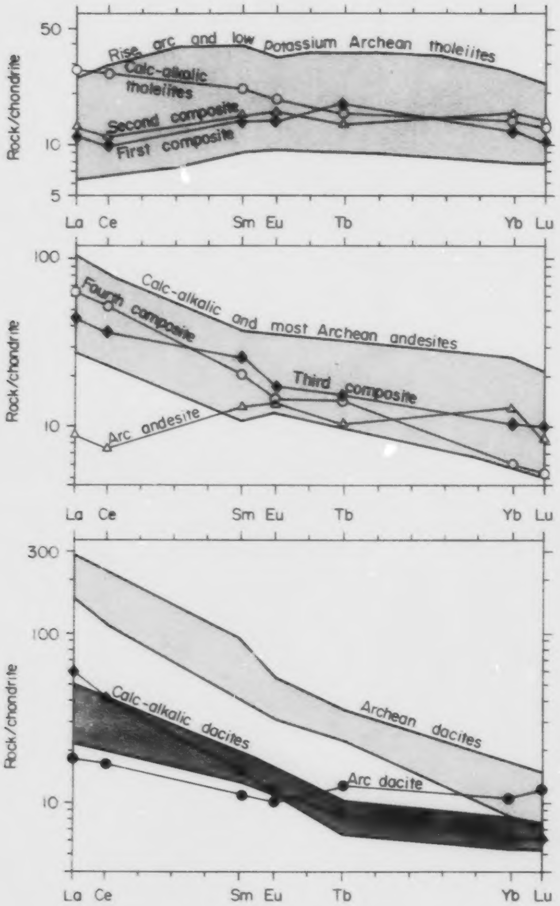


**Figure 19.62.** Chondrite-normalized rare earth element plots of calc-alkalic rocks from the Mazinaw Terrane. *Modified from* Condie and Moore (1977).

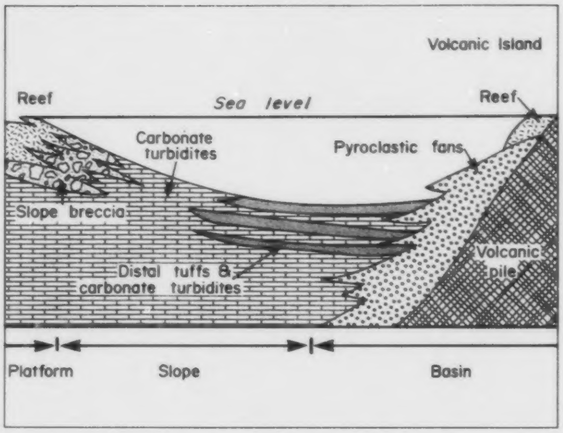


**Figure 19.63.** Sketch showing stratigraphic and facies relationships associated with mafic-felsic volcanic island complexes in the Central Metasedimentary Belt.

Hanmer et al. 1985). Another structural distinction between the Elzevir and Bancroft terranes is the abundance of thin sheets of metaplutonic rocks, probably thrust nappes, within the Bancroft Terrane, as opposed to more typical elongate or blob-like pluton patterns observed within the Elzevir Terrane. This latter structural definition would place most of the undersaturated rocks within the Elzevir Terrane but would place the boundary of the Bancroft Terrane immediately north of the Abinger Granite.

Another lithologic characteristic of the Bancroft Terrane is the presence of rocks of the *circa* 1340 million-year-old Dysart Suite *of* Lumbers et al. (1990) in the southern segment of the Bancroft Terrane between Haliburton and Norland (e.g., Redstone Lake Tonalite Gneiss, Dysart Tonalite Gneiss). Using the distribution of these bodies results in a boundary similar to that outlined by the distribution of thrusted plutonic masses.

Carlson et al. (1990) use the Bancroft shear zone as the Bancroft–Elzevir terrane boundary. Most workers have included the area between the Glamorgan Gneiss Complex and the Harvey–Cardiff Arch within the Bancroft Terrane (e.g., Davidson 1986a; Galway "Basin" *of* Haynes 1986), even though this area includes less intensely deformed rocks. In part, this is because the Tory Hill and Gooderham syenite belts lie within this area. Davidson et al. (1990, their Figure 2), however, show the Bancroft–Elzevir boundary as lying adjacent to the southeast margin of the Glamorgan Gneiss Complex, thereby excluding this region from the Bancroft Terrane. The boundary of Carlson et al. (1990) is used herein.

In summary, like the Sharbot Lake Terrane (Moore 1982; Easton 1988a), the Bancroft Terrane can be distinguished from the Elzevir Terrane by several lithologic criteria, and an approximate boundary defined, even though the Bancroft Terrane may not be fundamentally distinct from the Elzevir Terrane (i.e., the terranes may be parautochthonous *versus* allochthonous). This is in marked contrast to the Bancroft Terrane–Central Gneiss Belt boundary or the Elzevir–Mazinaw and Sharbot Lake-Frontenac terrane boundaries, which probably do mark the boundaries between allochthonous terranes accreted to one another during the Grenville Orogeny.

## Basement (?) to the Supracrustal Sequence

Various workers have suggested that the metasedimentary rocks of the Bancroft Terrane were deposited on gneisses of the Central Gneiss Belt (e.g., Schwerdtner and Lumbers 1980; Lumbers 1982; Lumbers et al. 1984; Carter 1984), and that a basal clastic sequence is present at the base of the Haliburton Group (cf., Lumbers et al. 1984). Within the Central Metasedimentary Belt Boundary Zone, however, rocks can locally be recognized as deformed equivalents of Central Gneiss Belt or Central Metasedimentary Belt rocks (Easton 1987d, 1987e, 1990a), and no compelling evidence for the presence of a basal clastic sequence to the Haliburton Group has yet been found.

The presence of basement is predicated solely on the basis of field relations, as there is no geochronology on proposed basement in the Bancroft Terrane and the adjacent

Fishog and Opeongo domains of the Central Gneiss Belt. If the Haliburton Group and its tectonically disrupted equivalents are time-correlative with the Hermon and Mayo groups in the Elzevir Terrane, then sedimentation in the Haliburton Group occurred between *circa* 1300 and 1250 Ma. This is some 40 to 50 million years younger than ages reported for the Redstone Lake Tonalite Gneiss and Dysart Tonalite Gneiss. If so, then these *circa* 1340 million-year-old tonalite gneisses could represent basement to the stratigraphic sequence, even if it is tectonically transported basement. It has been proposed that the Fishog Domain (Central Gneiss Belt) is structurally and lithologically equivalent to the Pine Island migmatites of the Go Home Domain, which Culshaw et al. (1991) have suggested is

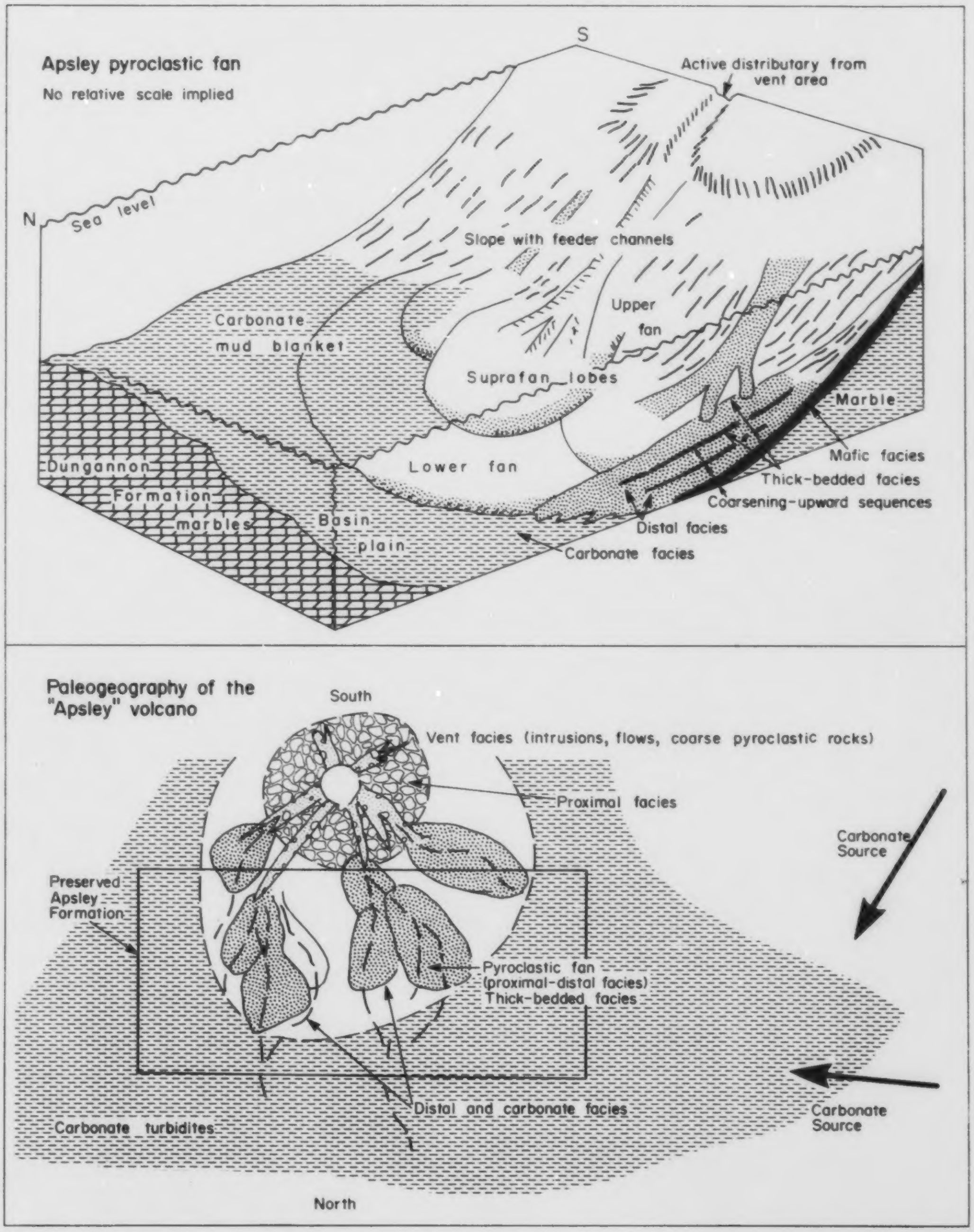

Apsley pyroclastic fan
No relative scale implied
S
N
Sea level
Active distributary from vent area
Slope with feeder channels
Upper fan
Carbonate mud blanket
Suprafan lobes
Marble
Mafic facies
Thick-bedded facies
Coarsening-upward sequences
Distal facies
Carbonate facies
Lower fan
Basin plain
Dungannon Formation marbles
Paleogeography of the "Apsley" volcano
South
Vent facies (intrusions, flows, coarse pyroclastic rocks)
Proximal facies
Carbonate Source
Preserved Apsley Formation
Pyroclastic fan (proximal-distal facies) Thick-bedded facies
Distal and carbonate facies
Carbonate turbidites
Carbonate Source
North

**Figure 19.64.** Sketch showing relationship of volcaniclastic rocks of the Apsley Formation and carbonate turbidites of the Dungannon Formation. Inset shows paleogeography of the "Apsley" volcano and the relative location of the presently preserved section.

younger than 1450 million years in age. If this is correct, then Central Gneiss Belt rocks adjacent to the Bancroft Terrane in the Minden–Haliburton area may be indistinguishable in age from the tonalite gneisses, and would confirm the suggestion of various workers (Lumbers 1982; Lumbers et al. 1984; Carter 1984) that the southwestern part of the Central Gneiss Belt was basement to the Haliburton Group, or that the Bancroft Terrane basement is indistinguishable in age from nearby Central Gneiss Belt rocks.

Lumbers et al. (1990) have suggested that the Dysart Suite tonalites intruded carbonate sediments in the area, and that therefore, sedimentation occurred prior to *circa* 1340 Ma, and that the supracrustal rocks in the Haliburton area are much older than those of the Grenville Supergroup (*sensu lato*). Lumbers et al. (1990) termed the supracrustal rocks the Haliburton sequence. Without additional age determinations on the sedimentary rocks themselves, these stratigraphic questions cannot be resolved.

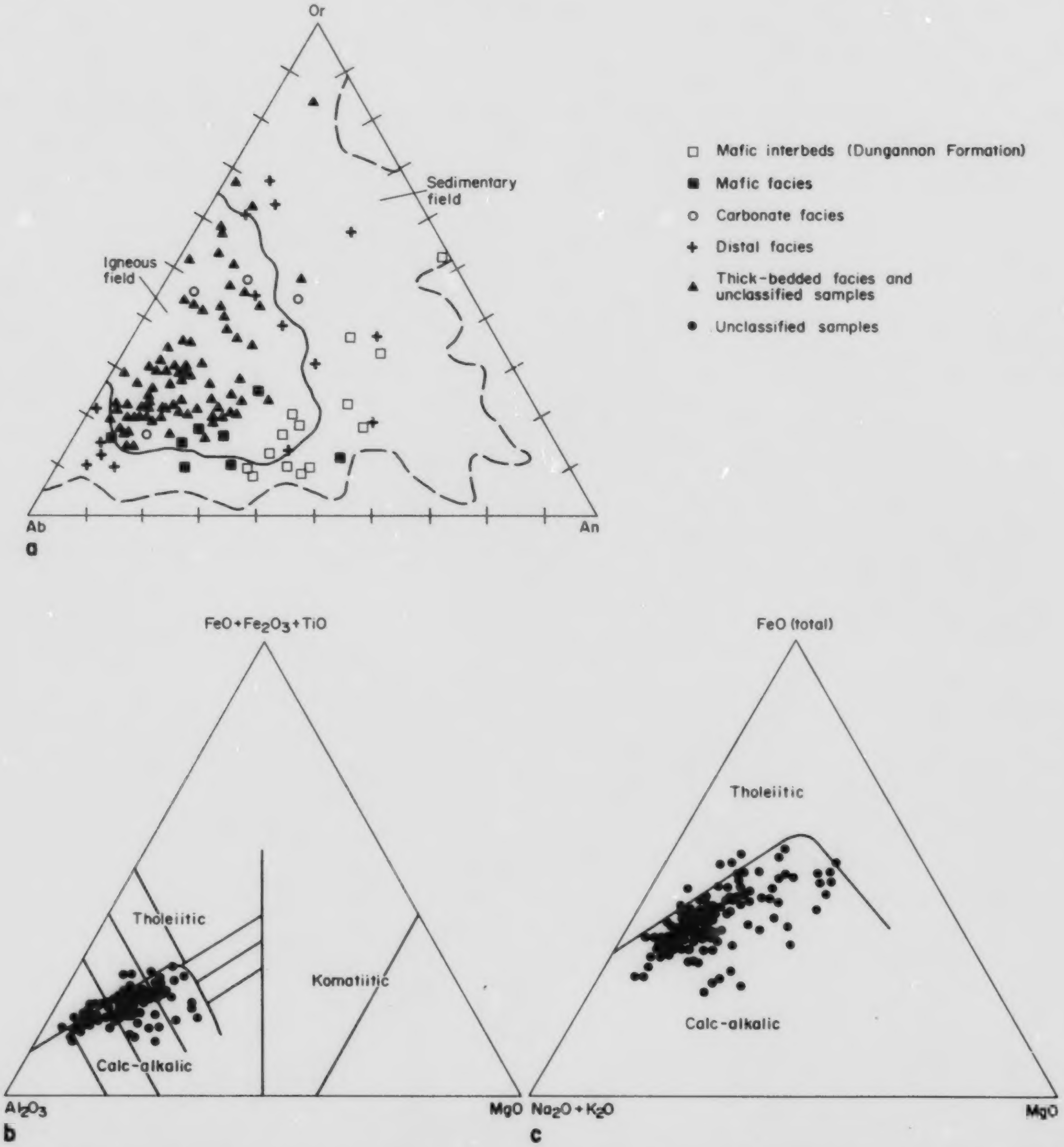


**Figure 19.65.** Geochemistry of the volcaniclastic Apsley Formation. **a)** Ab-Or-An plot *modified from* Shaw (1972). Sedimentary and igneous fields *from* Shaw (1972). Plot includes data *from* Easton, van de Kamp (1968), Shaw (1972) and Heaman (1985); dataset includes 137 major element analyses. Facies assignment of points based on field mapping. **b)** Jensen plot for Apsley Formation samples; dataset as for Figure 19.65a. **c)** AFM plot for Aplsey Formation samples; dataset as for Figure 19.65a.

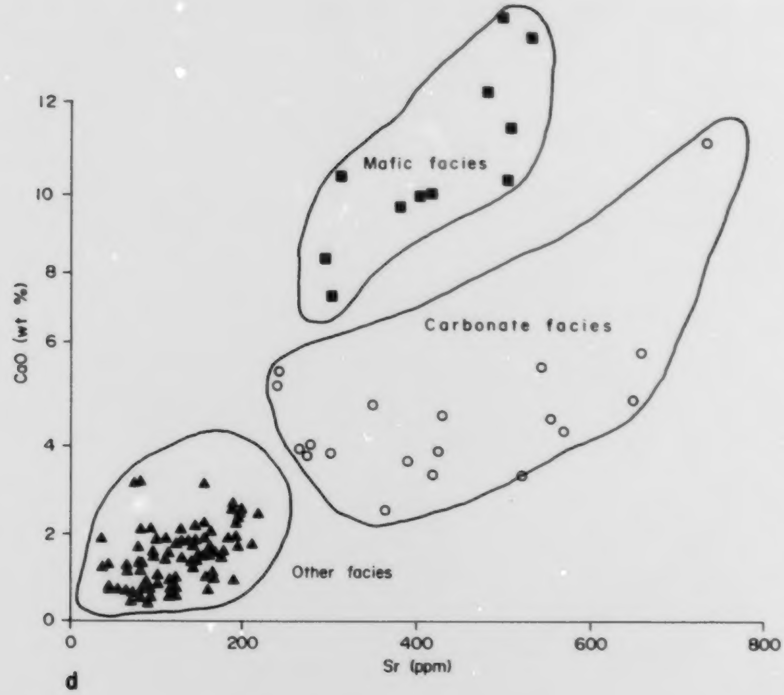


**Figure 19.65d.** Geochemistry of the volcaniclastic Apsley Formation. CaO-Sr plot showing variation between different facies of the Apsley Formation; dataset as Figure 19.65a.

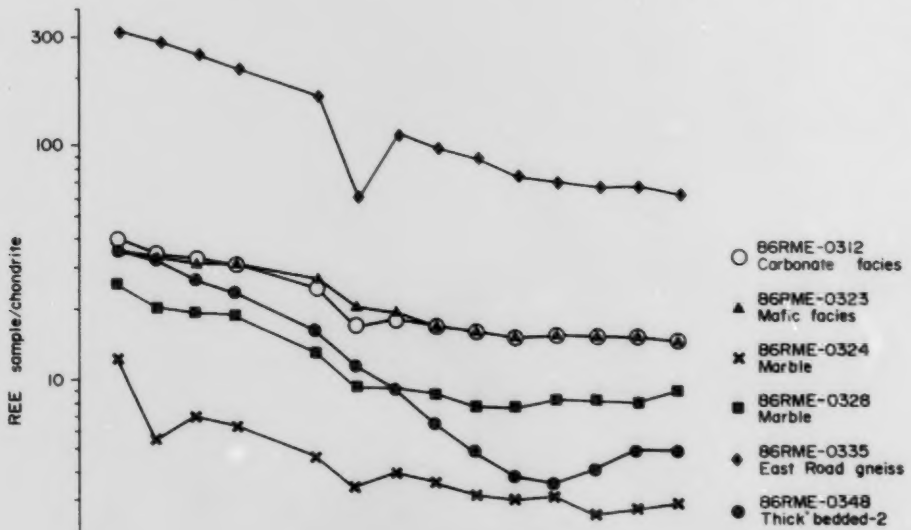


**Figure 19.65e.** Rare earth element geochemistry of Apsley Formation samples. Sample 86RME–312 is a marble with a pronounced cerium anomaly as would be expected from a marine sediment. Sample 86RME–324 is a marble interbedded with the Apsley Formation and shows a similar pattern to a dacitic volcaniclastic sample (86RME–348). Samples 86RME–312 and 86RME–323 are basaltic andesite samples. Dacite volcanic sample *of* Condie and Moore (1977) has a pattern identical to that of sample 86RME–348.

**Figure 19.66a.** Turbidites within the Mazinaw Terrane west of Plevna. Hammer handle is 30 cm long.

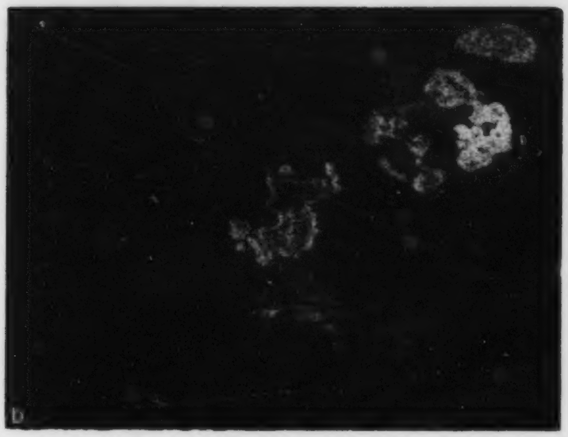


**Figure 19.66b.** Cross-bedded arenites, Belmont Lake, Belmont Domain. Knife is 9 cm long.

**Figure 19.66c.** Thin-bedded, volcanic-derived arenites, Belmont Lake, Belmont Domain. Knife is 9 cm long.

**Figure 19.66d.** Volcanic-derived wacke interbedded with thinly bedded carbonate turbidites of the Dungannon Formation, Belmont Domain, Highway 28 north of Apsley. Hammer handle is 30 cm long.

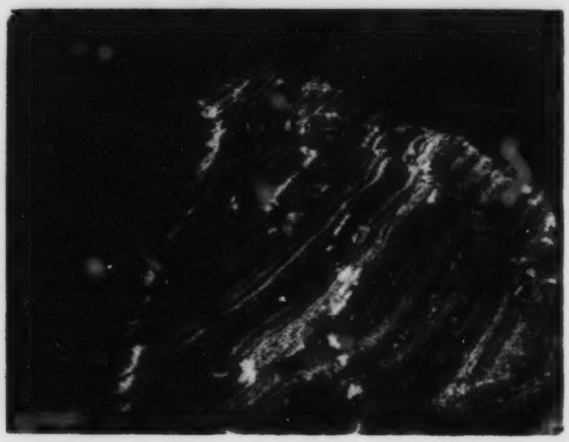


**Figure 19.66e.** Thinly bedded carbonate turbidites of the Dungannon Formation. Hammer handle is 30 cm long.

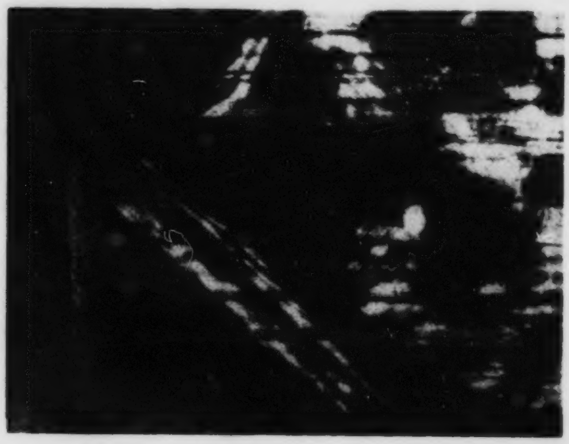


**Figure 19.66f.** Close-up of carbonate turbidites within the Sharbot Lake Terrane. Hammer handle is 30 cm long.

## Haliburton Group

### INTRODUCTION

Formal stratigraphy exists locally in the southern part of the Bancroft Terrane (*see* Table 19.9). It is in this area, where formations can locally be defined, that the term Haliburton Group is best applied (e.g., Bright 1977; *see* Table 19.9).

In the western parts of the Bancroft Terrane, stratigraphy is chaotic, with local remnants of Haliburton Group stratigraphy mapped as a ghost stratigraphy. The western and northern parts of the Bancroft Terrane are best described as structural complexes (NACSN 1983) of tectonically disrupted Haliburton Group strata. The Denna Lake Structural Complex represents one such unit (Easton 1983, 1990a; *see* Figures 19.72, 19.73 and 19.74).

The discussion below focusses on lithologic associations, which are informal units comparable to groups in scale and scope. They are discussed in rough stratigraphic order from oldest to youngest.

### DOLOMITE-QUARTZITE ASSOCIATION

The lowermost part of the sequence consists of lenses of relatively pure, quartz arenites and feldspathic litharenites, with minor quartzose, polymictic conglomerates containing quartz arenite and granitoid cobbles; and relatively massive, thickly layered dolomitic marbles, locally stromatolitic (Easton 1987a, 1987e). Graded bedding and cross-bedding are present within the quartz arenites and litharenites, as are magnetite-rich layers, possibly heavy mineral lags. This sequence occurs within the Denna Lake Structural Complex and southeast of the Glamorgan Gneiss Complex where an undisrupted stratigraphic sequence is preserved. In the latter area, Bright (1976, 1977, 1987a) separated the association into 2 formations: the Salerno Lake Formation (quartz arenites and related quartzofeldspathic gneisses) and dolomitic marbles of the Dungannon Formation. The marbles can be further subdivided. Dolomites dominate low in the stratigraphy; the abundance of interlayered calcitic marble beds or layers increasing with stratigraphic height until calcite marbles predominate. The proportion of siliceous clastic metasedimentary rocks also decreases with stratigraphic height. Dolomitic marbles are commonly tremolitic low in the sequence where stromatolites (generally algal-mat type) are present (Easton 1987a). Easton (1987a) envisaged a near-shore environment for this sequence, with the quartzose sediments being deposited in channels within a shallow-water carbonate sequence. Dolomitization may be due to diagenesis, and not primary dolomite deposition. A similar association is present in the Denbigh–Matawatchan area (cf., Themistocleous 1981a, 1981b).

### AMPHIBOLITE, VOLCANIC AND VOLCANICLASTIC ASSOCIATION

Metavolcanic rocks are a minor component in the Bancroft Terrane and are mostly found within structural complexes. Near Crystal Lake, however, metavolcanic rocks within the Haliburton Group include dacitic tuffs and lapilli tuffs, as well as metasiltstones and metalitharenites of dominantly volcanic provenance (Easton 1987a). Also present are "para-amphibolites" which probably represent volcanic-derived wackes. The distribution of metavolcanic rocks within the Bancroft Terrane, and the rock types present, are consistent with them being distal components of volcanism taking place east and south of the Bancroft Terrane. These units probably occur in the structural (and stratigraphic?) lower part of the Haliburton Group (Easton 1987a, 1987e; *see* Figure 19.72b).

Orthoamphibolite is abundant in the Denna Lake Structural Complex and within the Gelert Lithodeme of the Glamorgan Gneiss Complex (Easton 1987e). These are dark, featureless hornblende-andesine rocks, commonly associated with rusty paragneisses and dolomitic marbles. Chemically they are alkalic, however, this could reflect local fenitization, not a primary chemical signature, although rare earth element data from these rocks shows an alkalic basalt signature (Easton 1987e). No primary textures are present in these rocks. Their limited distribution and chemistry distinguish these rocks from the major volcanic sequences of the Elzevir Terrane.

Metavolcanic rocks are a minor component in Ashby, Denbigh and Lyndoch townships. Most are basaltic (probably tholeiites), with rare pillows (Evans 1964). Far more abundant are garnetiferous para-amphibolites intercalated with wackes and pelites, commonly sillimanite-bearing, and calcitic marbles.

### PARAGNEISS ASSOCIATION

Easton (1987c, 1987e) noted an association between yttrium-strontium-zirconium-neodymium-enriched rusty paragneisses and calcitic marbles, alkalic amphibolites, diopside-plagioclase rocks and chemically unusual anorthosites, such as the Allsaw Anorthosite, within the Denna Lake Structural Complex in the Haliburton area. Locally, associated with this sequence are cordierite-gedrite ± sillimanite-bearing rocks interpreted as paleoregoliths (Breaks and Thivierge 1985; Culshaw 1986) or restites (Lal and Moorhouse 1969). Easton (1987c, 1987e) suggested that this association may represent metamorphism of a nepheline syenite-carbonatite complex and associated phonolitic volcaniclastic detritus and related fenites. However, they could be metasomatized supracrustal rocks related to the fenite-carbonatite-pegmatite suite (Lumbers et al. 1990).

Cordierite-gedrite±sillimanite-bearing rocks occur near the Bancroft Terrane–CMBBZ boundary (Figure 19.75), and several workers have suggested that these rocks mark the basal thrust of the Haliburton and Pembroke thrust "stacks" (Hanmer 1988; McEachern 1990).

### INTERBEDDED CALCITE MARBLE AND QUARTZOSE SEDIMENT ASSOCIATION

Calcitic marbles, the most abundant rock type in the succession, are generally massive, or layered on a decametre to metre scale. Where the calcitic marbles are in proximity to the dolomite-quartzite association, thin, centimetre-scale interbeds of quartzose arenite and litharenite are intercalated with the calcitic marbles on a centimetre to decametre scale. Calc-silicate layers are also found throughout the calcite marble sequence, but again, are most common where the calcite marbles overlie the quartzite-dolomite or metavolcanic associations (Easton 1985b, 1987a). In the Tory Hill area, southeast of the Glamorgan Gneiss Complex, Bright (1976, 1977, 1987a) introduced the term Tory Hill Formation for this sequence of interbedded calcite marbles, calc-silicates and quartzofeldspathic metasedimentary rocks. In the Denbigh-Matawatchan area, this facies association is widespread, although no formations have been proposed.

### DENNA LAKE STRUCTURAL COMPLEX

A structural complex is defined as follows (NACSN 1983):

> In some terranes, tectonic processes (e.g. shearing, faulting) have produced heterogeneous mixtures of disrupted bodies of rock in which some individual components are too small to be mapped. Where there is no doubt that the mixing or tectonic disruption is due to tectonic processes, such a mixture is a structural complex, whether it consists of two or more classes or rock, or a single class only.

Easton (1983, 1990a) introduced the term the Denna Lake Structural Complex to refer to a predominantly marble tectonic breccia zone in the Minden–Haliburton area lying between the Glamorgan Gneiss Complex and the Central Metasedimentary Belt Boundary Zone (*see* Figures 19.72, 19.73 and 19.74). Many of the stratigraphic units recognized south of the Glamorgan Gneiss Complex (i.e., various associations listed above), as well as the plutonic rocks intruding them, are found in a state of tectonic disruption within the Denna Lake Structural Complex, producing a "ghost stratigraphy" (Easton 1985b, 1986b, 1990a; *see* Figure 19.72). Mineral deposit types show an association with this ghost stratigraphy (cf., Easton 1985b, 1990a; Figure 19.76). Figure 19.77 compares the map pattern of the area underlain by the Denna Lake Structural Complex and the area southeast of the Glamorgan Gneiss Complex. The structural complex concept is appropriate for 1:50 000 scale mapping of these marble breccia zones, and the concept can probably be applied to parts of the Bancroft Terrane.

### FACIES VARIATION ACROSS THE BANCROFT TERRANE AND DEPOSITIONAL ENVIRONMENTS

In the southwestern part of the Bancroft Terrane, there is a west to east change from a near-shore facies (dolomite-quartzite association) to a far shore-slope facies (distal volcaniclastics, interbedded clastics-calcite marbles). This transition may have been transgressive and was followed by a calcitic marble succession. Abundant arenites, stromatolites and lack of carbonate turbidites are all consistent with a platformal, shallow-water environment within the southern part of the Bancroft Terrane. Similar rock types are abundant in the northeastern Bancroft Terrane, and a similar facies change and depositional setting were probably present.

### CORRELATION WITH THE GRENVILLE SUPERGROUP IN THE ELZEVIR TERRANE

As noted previously in this chapter, the Haliburton Group may be older than *circa* 1340 Ma, considerably older than the Grenville Supergroup. If instead the Haliburton Group were deposited between roughly 1300 and 1250 Ma, then it may be time-correlative with the Grenville Supergroup in the Elzevir Terrane. In this case, the dolomite-quartzite and metavolcanic associations may be correlative with the Hermon or Cardiff groups, with the latter representing distal equivalents of either group.

## Plutonism

### EARLY TONALITE (DYSART) SUITE (ca. 1340 Ma)

The Dysart, Fishtail Lake and Elephant Lake tonalitic gneisses represent the largest plutons of this suite, which are located only along the northwestern margin of the Bancroft Terrane in the Haliburton–Bancroft area. Zircons from this suite are in the 1370 to 1340 million year age range (Lumbers et al. 1990), similar in age to the nearby Redstone Lake Tonalite Gneiss in the CMBBZ ($1344^{+93}_{-32}$ Ma, van Breemen and Hanmer 1986).

Hornblende tonalite and granodiorite dominate the suite. Metamorphic clinopyroxene is locally present, and dikes of aplite and granite pegmatite are common (Lumbers et al. 1990). Analyses of rocks from this suite are given in Lumbers et al. (1990) and Culshaw (1986).

Lumbers et al. (1990) compared the chemistry of the Early Tonalite Suite (Dysart Suite) and younger tonalites such as the Elzevir and Weslemekoon tonalites. The early tonalites have lower $K_2O/MgO$ ratios and all lie along or below the trondhjemite trend (Arth 1979), whereas the younger tonalites show calc-alkalic and trondhjemitic trends, in part, because of the greater range in $K_2O$ contents in the younger tonalites. Lumbers et al. (1990) note the early tonalites are mainly hornblende-bearing, low-alumina types, whereas the younger tonalites are biotite-bearing high-alumina types (Arth 1979).

### NEPHELINE SYENITE SUITE (ca. 1290 TO 1250 Ma)

The nepheline syenite suite includes the nepheline gneisses of the Haliburton–Bancroft region, well known to petrologists through the classic work of Adams and Barlow (1910), Gummer and Burr (1943), von Eckermann (1942) and Tilley and Gittins (1961), among others. The suite is most extensive in the northern part of the Bancroft Terrane, forming a discontinuous zone up to 15 km wide (Figure 19.78).

Rocks of the suite are considerably deformed and recrystallized; many early studies concluded that the rocks were metamorphic or metasomatic in origin. The metasomatic models commonly refer to a process of "nephelinization" of country rocks. However, a few plutons (Trooper Lake–Laronde Creek, Miller 1983; Conroy Marsh, Sebastopol and Highland Lake, Lumbers 1982; Body "B" in Easton 1987a) contain primary igneous textures and mineralogy.

Many plutons are elongate in shape, a feature early workers attributed to metasomatism of sedimentary rock types, but which have now been demonstrated to reflect original emplacement as sills (cf., Easton 1987a; Lumbers et al. 1990). Some workers have suggested that parts of the sequence may have been volcanic in origin (Bright 1987a; Easton 1987a), however, there is no definitive evidence supporting this interpretation, and an intrusive origin best fits the available data.

According to Lumbers et al. (1990), the suite is dominated by miaskitic nepheline-bearing rocks, mainly nepheline syenite with lesser malignite, urtite and feldspathic urtite, ijolite and feldspathic ijolite, nepheline diorite and gabbro, nepheline-normative diorite and gabbro and albite syenite (Miller 1983, 1984; Lumbers et al. 1990). Nepheline diorite and gabbro are rare, but some bodies, such as the Sebastopol and Highland Lake intrusions, consist mainly of nepheline-normative diorite and gabbro (Lumbers et al. 1990). Lumbers et al. (1990) distinguish 3 varieties of nepheline syenite: 1) oligoclase nepheline syenite containing 0 to 10% potassium feldspar; 2) albite nepheline syenite with potassium feldspar making up 10 to 30% of the total feldspar component; and 3) potassium feldspar nepheline syenite with potassium feldspar accounting for over 30% of the total feldspar. All 3 types are commonly present in any individual body.

Average chemical analyses for the major rock types of the suite are available in Tilley and Gittins (1961), Armstrong and Gittins (1968) and Lumbers et al. (1990). Rocks of the suite are similar to nepheline-bearing rocks associated with carbonatite-nephelinite complexes in rift systems, such as the East African rift (Miller 1983).

Altered zircons in a gneissic nepheline syenite give a minimum age of 1219 Ma for the suite (Miller 1983; Lumbers et al. 1990). Lumbers et al. (1990) showed that the nepheline syenites are older than the *circa* 1270 million-year-old Weslemkoon Tonalite and the *circa* 1240 million-year-old Methuen Suite, but younger than the *circa* 1290 million-year-old tholeiitic volcanic rocks. Lumbers et al. (1990) therefore estimated the age range of the suite as 1290 to 1250 Ma.

## ANORTHOSITE SUITE (1280 TO 1250 Ma)

Rocks of this suite form small plutons spatially associated with the nepheline syenite suite (e.g., Lumbers 1982; Lumbers et al. 1990), or thrust sheets within the Denna Lake Structural Complex (e.g., Allsaw Anorthosite, Easton 1987e, 1990b; Farquahar Lake complex, Culshaw 1986) or the Central Metasedimentary Belt Boundary Zone (e.g., Percy Lake meta-anorthosite, Marmont et al. 1988). Most bodies consist of anorthosite and gabbroic anorthosite. Some bodies contain gabbroic and dioritic phases, generally near the margins of the bodies.

The Allsaw Anorthosite is a tectonic sheet surrounded by marble breccia of the Denna Lake Structural Complex (Easton 1987e) and represents a remnant of a more extensive body. The Farquhar Lake complex is a series of dismembered slices of metapyroxenite and metagabbro to meta-anorthosite (Culshaw 1986). These bodies are geochemically distinct from other Grenville meta-anorthosites in Ontario (*see* Figures 19.22 and 19.23; Easton 1990b). Both are characterized by low abundances of iron, magnesium, titanium, phosphorus, zirconium, yttrium and vanadium and by normative nepheline contents of 1 to 15%, and will be referred to as low ferromagnesian meta-anorthosites. The REE pattern of meta-anorthosites from the Allsaw Anorthosite is unlike that of massif-type meta-anorthosites in the Central Gneiss Belt (*see* Figure 19.23), but is very similar to the REE pattern of some metamonzodiorite bodies of the Gooderham Syenite Belt, suggesting a possible genetic relationship (Lumbers 1976b, 1982; Easton 1990b; Lumbers et al. 1990).

Taylor (1969) reported $^{18}$O values of 8.7 and 8.3 from whole-rock and plagioclase samples from the Allsaw Anorthosite, respectively. These values are similar to those found in other metamorphosed Grenville Province anorthosites.

## OLDER GABBRO SUITE (GREATER THAN 1270 Ma)

Also present within the Bancroft Terrane are several large metagabbroic complexes containing minor gabbroic anorthosite and anorthositic gabbro, notably the Boulter, Mallard Lake, Raglan Hills, O'briens Hill and Quadeville bodies (*see* Figure 19.75). The first 3 bodies are thrust sheets referred to as the Raglan gabbros. Insufficient work has been done on the geochemistry of these bodies to determine if they are related to the Killer Creek or the Lavant suites. The abundance of hornblendite and anorthositic phases, paucity of late diorite and tonalite, and occurrence of copper-nickel sulphide mineralization in these bodies suggest an affinity to the Killer Creek Suite. Recent U-Pb zircon geochronology gives ages of 1231±2, 1227±5 and 1246±3 million years, for the Chenaux, Raglan–Boulter–Mallard Lake, and Trooper Lake gabbroic intrusions, respectively (McEachern et al. 1992); suggesting that they are associated with the Lavant Suite. The lithologic similarity between parts of the Lavant Suite and the Killer Creek Suite may reflect similar tectonic settings of magma generation for the 2 suites, but separated in time.

## TONALITE (ELZEVIR) SUITE (1280 TO 1270 Ma)

The Glamorgan Gneiss Complex (1987e, 1990a) is the largest body of this suite in the Bancroft Terrane. Other bodies include the Union Lake Granodiorite (Easton 1987a; Wu 1984), the Boulter Tonalite, the Griffith, Mud Lake and Hurds Lake tonalites and the Slate Hills Complex. A suite of similar rocks is common within the Elzevir Terrane.

The Glamorgan Gneiss Complex consists of 3 dominant rock types: 1) a sodic suite (tonalite, with minor trondhjemite and granodiorite); 2) a potassic suite (syenogranite); and 3) amphibolite, slightly alkalic in composition and possibly volcanic in origin (Easton 1987a, 1987e, 1990a). The tonalitic rocks comprise the core of the complex, with the granitic rocks rimming the complex and are locally infolded into the core of the complex together with the amphibolites. Biotite and hornblende tonalites are present in the Glamorgan Gneiss Complex. Only the sodic suite

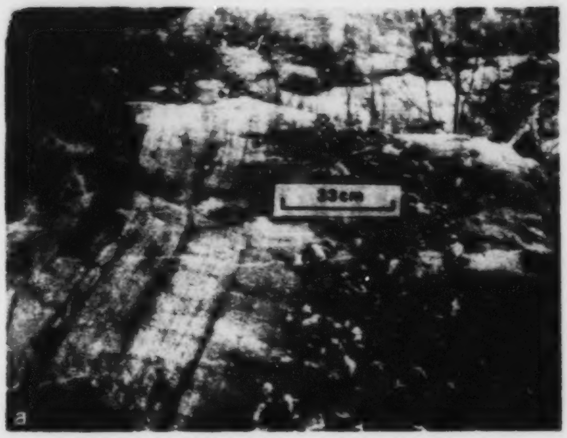


**Figure 19.67a.** Thinly layered coarse-grained calcite marbles (left) and calcite marble with quartz pods (right) from the Omega Blue marble quarry near Tatlock. Marbles are preserved at middle amphibolite facies. Scale bar is 33 cm long.

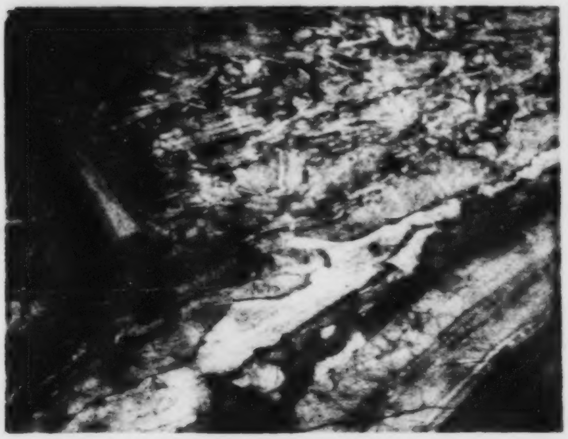


**Figure 19.67b.** Carbonate chip breccia bed within thinly layered calcitic marbles, Highway 62, Belmont Domain. Hammer handle is 30 cm long.

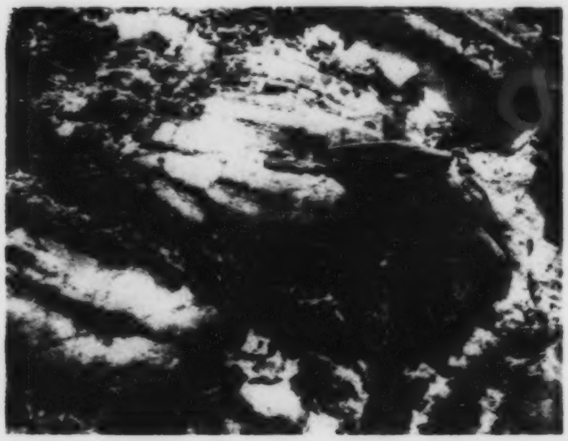


**Figure 19.67c.** Algal-laminite stromatolites in dolomitic marbles, Canada Talc Mine, Madoc, Belmont Domain. Hammer handle is 30 cm long.

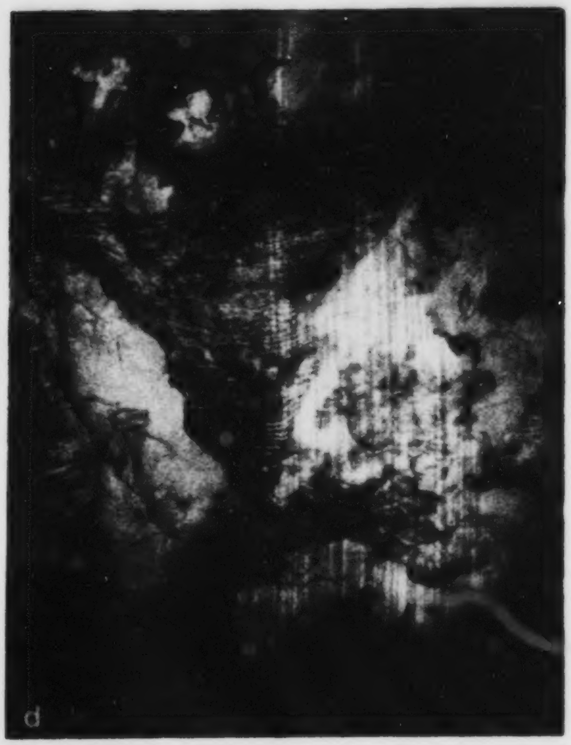


**Figure 19.67d.** Columnar stromatolites within dolomite, Big Island, Belmont Lake, Belmont Domain. Hammer head is 15 cm across.

member is comparable to the Elzevir Suite. The syenogranite may be comparable to the Methuen Suite.

Geochemical studies on the Glamorgan Gneiss Complex include Armstong and Gittins (1968), Chesworth (1970), Pride (1983) and Easton (1987a, 1987e). Rare earth element data on the sodic phases of the Glamorgan Gneiss Complex (Easton 1987e) is compared with the Elzevir Tonalite (Pride and Moore 1983) in Figure 19.79b. Major element geochemistry of both bodies is similar and characteristic of high-alumina trondhjemites from continental regions (Arth 1979). Trace and rare earth element data (*see* Figure 19.79) indicate that the Glamorgan Gneiss Complex has a calc-alkalic signature and was likely derived from melting of a crustal source of granodioritic composition, or crustal contamination of the mantle-derived magma. These geochemical differences may be related to the location of the Glamorgan Gneiss Complex in the Bancroft Terrane in an area where slightly older crust may be present.

## DIORITE-GABBRO (LAVANT) SUITE
## (1250 TO 1240 Ma)

Rocks of this suite are less abundant in the Bancroft Terrane than in the Elzevir Terrane. Both diorite and gabbro plutons occur, including the Salerno Lake Gabbro (Easton 1987a), the Bark Lake Diorite (Easton 1987e), the Minnecock Lake Diorite and several small bodies near Furnace Falls (Easton 1987a). As previously noted, the Chenaux, Raglan–Boulter–Mallard Lake, Trooper Lake and Faraday gabbroic

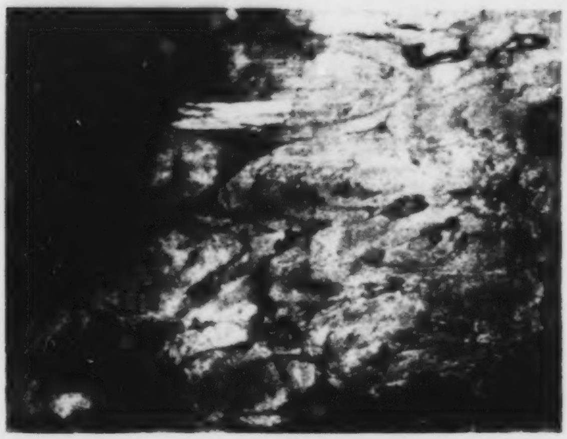


**Figure 19.67e.** Cross section of domal stromatolites within dolomitic marbles, L'Amable. Lens cap (upper right) is 5 cm across.

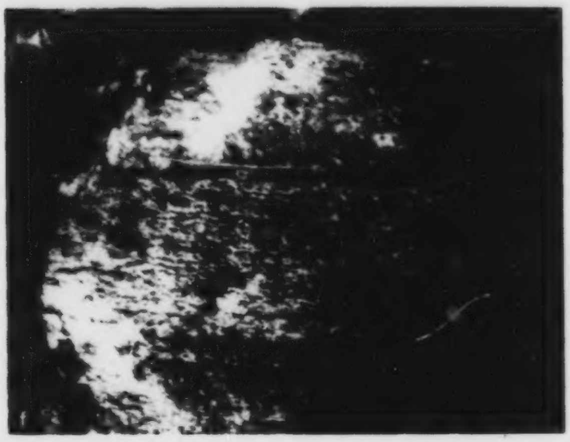


**Figure 19.67f.** Algal-laminite stromatolites in dolomitic marbles preserved at upper amphibolite facies, Mazinaw Terrane, west of Clyde Forks. Hammer handle is 30 cm long.

intrusions likely belong to the Lavant Suite as well. In the south, at least, the 2 rock types are present in distinct bodies. The Bark Lake Diorite intrudes the Glamorgan Gneiss Complex forming intrusion breccias along its margins (Easton 1987e).

Diorites have colour indices of 20 to 35 and are hornblende-oligoclase and quartz-biotite-andesine rocks. Igneous textures are preserved in the cores of the bodies, despite recrystallization during metamorphism. Gabbros have a higher colour index and contain clinopyroxene-hornblende-oligoclase and locally labradorite. Scapolite is a common alteration mineral in the margins of the bodies, particularly where gabbros intrude marbles. An alaskitic syenogranite is associated with the Salerno Lake Gabbro (Easton 1987a), but contact relations could not be deduced.

Magnetite skarns are commonly associated with gabbros intruding marbles. Sulphide skarns, mainly pyrite-pyrrhotite-chalcopyrite, with local enrichments of zinc and cobalt are associated with diorites intruded into marbles (Easton 1987a). Chemical analyses of diorite and gabbros are given in Easton (1987a, 1987e), Armstrong and Gittins (1968) and Bright (1987a).

Several large, alkalic gabbro bodies, including the Faraday, Hadlington, Glamorgan and Monmouth gabbros, are spatially associated with the nepheline syenite suite and the fenite-carbonatite-pegmatite suite, and have higher alkalic contents, particularly $K_2O$ (greater than 0.75 wt % compared to less than 0.50 wt %), than the gabbros noted above. Many of the gabbros contain corona textures (Bright 1987a). In some instances, such as the Trooper Lake phase of the Glamorgan Gabbro which is nepheline bearing (Bright 1987a), these intrusions may simply be mafic end-members of the nepheline syenite suite. However, other parts of the body, more distal to the nepheline syenite suite bodies, are olivine normative (Bright 1987a). Other gabbro bodies, such as the Faraday Gabbro, may be alkali metasomatized members of the Lavant Suite.

## ALASKITE (METHUEN) SUITE (1250 TO 1240 Ma)

Generally small plutons of this suite occur throughout the Bancroft Terrane with some probable intermixing with other plutonic rocks as well, for example, the syenogranites of the Glamorgan Gneiss Complex. As noted by Lumbers et al. (1990), many alaskite suite plutons in the northern Bancroft Terrane have been fenitized during the emplacement of the fenite-carbonatite-granite pegmatite suite.

## FENITE AND CARBONATITE SUITE (1070 TO 1040 Ma) AND PEGMATITE SUITE (1050 TO 1030 Ma)

Lumbers et al. (1990) have documented a suite of carbonatitic rocks and associated fenites, ranging in age from 1070 to 1040 Ma (Lumbers et al. 1990), the latter characterized by patches and veinlets of alkali pyroxene and amphibole (Figure 19.80), in a band running the length of the Bancroft Terrane paralleling the CMBBZ. The fenitizing fluids introduced sodium, potassium, calcium, iron, titanium, phosphorus, fluorine, chlorine, uranium, thorium, rare earth elements, zirconium, barium and molybdenum into the country rocks (Lumbers et al. 1990). The carbonatite suite includes the so-called calcite "vein dikes" of the Bancroft region, the origin of which has long been controversial (*see* next paragraph).

Associated with these rocks are granitoid pegmatites, some of which are uranium-, molybdenum- and rare element-bearing. Although late pegmatites occur throughout the Central Metasedimentary Belt, they are most abundant in the general vicinity of the fenite-carbonatite belt, particularly the uranium, molybdenum and rare-element pegmatites. Lentz (1991b) describes the petrology of the fenites and their associated pegmatites and carbonatites.

Several hypotheses have been proposed for the origin of these pegmatites, skarns and veins, as summarized by Lentz (1991b).

### Pegmatites

It has been suggested that the radioactive pegmatites formed by: 1) *in situ* anatexis of a uraniferous protolith (Heinrich 1959; Robinson 1959; Allen 1971; Tremblay 1974; Ford 1982, 1983) or 2) fractionation from a granitic melt derived at depth(?) possibly associated with granite bodies (Fowler and Doig 1983; Lentz and Kretz 1989; Cerny 1990).

### Uranium-Molybdenum Skarns

Many models have been proposed for the origin of the uranium-molybdenum skarns, including: 1) a metamorphic-hydrothermal origin for the molybdenite-bearing skarns (Karvinen 1973; Carter et al. 1980), 2) a syngenetic origin for some uraniferous skarns (Masson and Gordon 1981; Haynes 1986) and 3) a magmatic-hydrothermal origin for these skarns (uranium, molybdenum and others; Wilson 1921; Landes 1938; Currie 1951; Lang 1952; Satterly 1957; Vokes 1963; Masson and Gordon 1981; Storey and Vos 1981b; Lentz 1991a).

### Calcite Veins

The uranium-, molybdenum- and REE-bearing fluorite-apatite-calcite veins ("vein-dikes" *of* Ellsworth 1932) originated by either: 1) metamorphic processes (Lafontaine 1980), 2) the pegmatitic stage of mafic magmas (deSchmid 1920; Spence and Carnochan 1930), 3) carbonatites (Heinrich 1966; Mungall 1989; Lumbers et al. 1990) or 4) hydrothermally remobilized carbonate veins related to skarn formation (Wilson 1921; Landes 1938; Currie 1951; Rowe 1952; Powell 1965; Lentz 1991a).

Lentz (1991a, 1991b) favours a magmatic-hydrothermal origin for the skarns and veins related to the emplacement of chemically evolved granitic pegmatites at intermediate crustal levels. The magmatic-hydrothermal fluids were derived from the crystallizing granitic pegmatites at volatile-phase saturation and were responsible for skarn formation. The chemically modified hydrothermal fluid produces "fenite-like" alteration due to calcium, magnesium, alkali and iron metasomatism during deposition of calcite, apatite and fluorite in veins along older structures. Lentz (1991b) suggested that this occurred in the abyssal environment (at roughly 500°C and 4 kb) based on thermochronologic studies (Berger and York 1981a, 1981b). In contrast, Lumbers et al. (1990) relate all pegmatites in the Central Metasedimentary Belt and the zone of carbonatite and fenites in the Bancroft Terrane to a regional metasomatic event generated by high-temperature alkalic fluids which rose through the crust along dilatant fractures.

The carbonatites are generally small, elongate bodies, generally less than 1 m wide, but a few, such as the Silver Crater carbonatite, crop out as circular masses. Zircon ages from the carbonatites and fenites from the zone (Lumbers et al. 1990) lie in the range 1070 to 1038 Ma, and ages from the fenites span a similar range from 1072 to 1044 Ma (Lumbers et al. 1990; Mezger et al. 1991). The Sullivan Island Carbonatite Complex in the Ottawa River south of Pembroke has been zircon dated at *circa* 1050 Ma (Lumbers 1982) and is therefore considered part of this suite, and unrelated to Ottawa–Bonnechere Graben alkalic magmatism.

The localization of pegmatite injection, fenitization and carbonatite emplacement near the Bancroft Terrane/Central Metasedimentary Belt Boundary Zone may be structurally controlled. As noted in the geophysics section of this chapter, the CMBBZ extends to the base of the crust, the only such deep crustal structure noted in the CMB other than the Ottawa–Bonnechere Graben.

## Metamorphism

### TIMING OF METAMORPHISM

The Bancroft Terrane is preserved at amphibolite facies, ranging from lower amphibolite facies in the southeastern part of the terrane, increasing to upper amphibolite in proximity to the CMBBZ. P-T conditions are presented in Table 19.2. Three periods of metamorphism may be present. A contact metamorphism at *circa* 1250 Ma associated with nepheline-syenite, diorite-gabbro and alaskitic plutonism, and a regional metamorphism at *circa* 1180 to 1160 Ma (Mezger et al. 1990) coincident with regional metamorphism across the Central Metasedimentary Belt. Isotopic resetting associated with fenite-carbonatite-pegmatite magmatism has occurred locally, and it is probably this event that Mezger et al. (1990) attributed to a 1045 to 1020 million-year-old metamorphism. Rb-Sr whole-rock ages of 1187±56 Ma and 1164±26 Ma from nepheline syenite suite intrusions (Miller and Gittins 1982) may also record this regional metamorphic event. $^{40}Ar/^{39}Ar$ ages from hornblende and biotite from the Bancroft Terrane are in the 990 to 970 and 950 to 940 million year age range

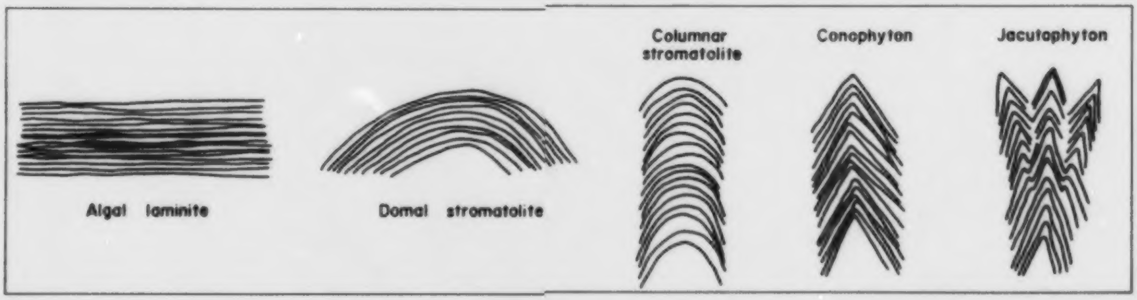


**Figure 19.68.** Stromatolite types described from marbles within the Central Metasedimentary Belt (*modified from* Bartlett and deKemp 1987).

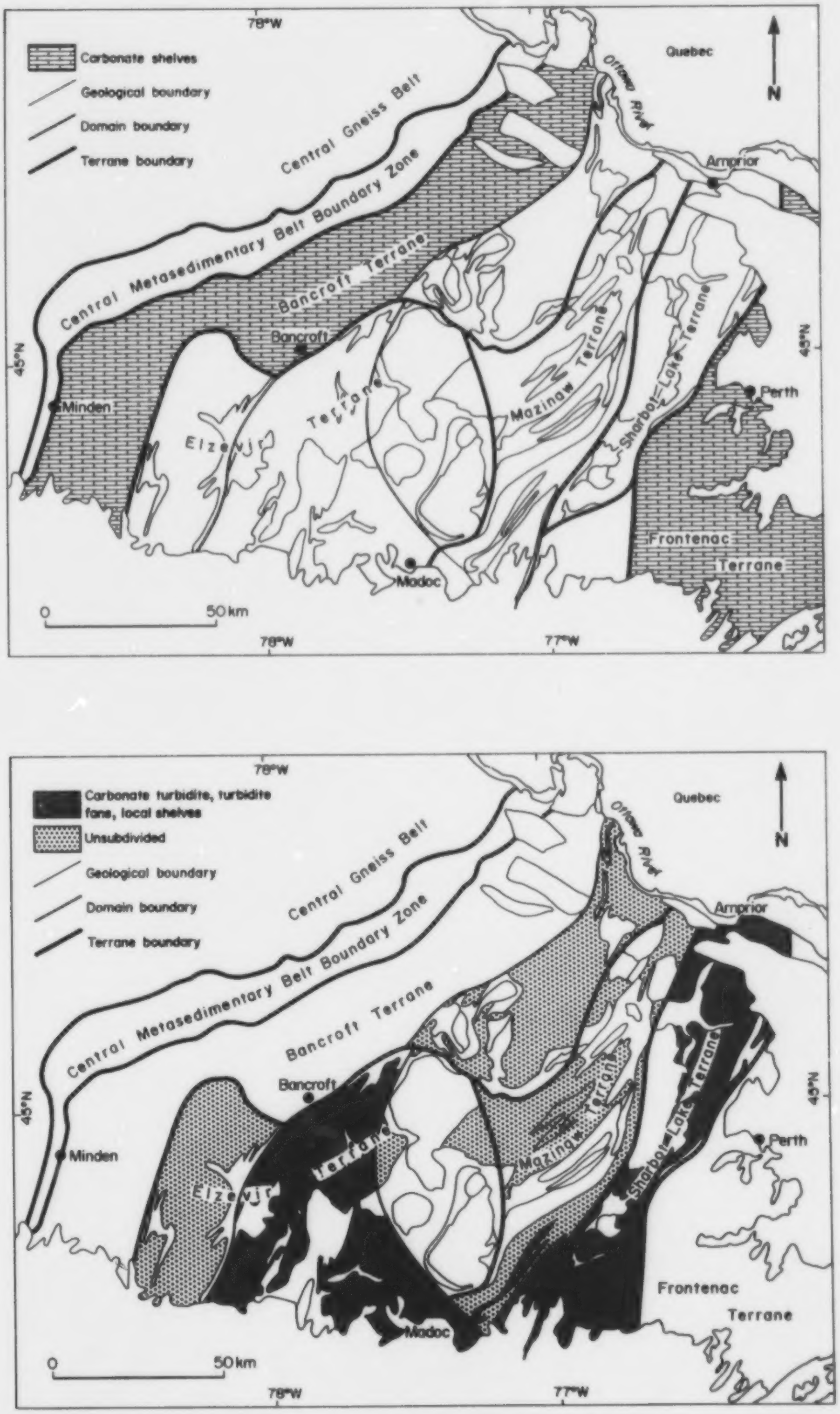
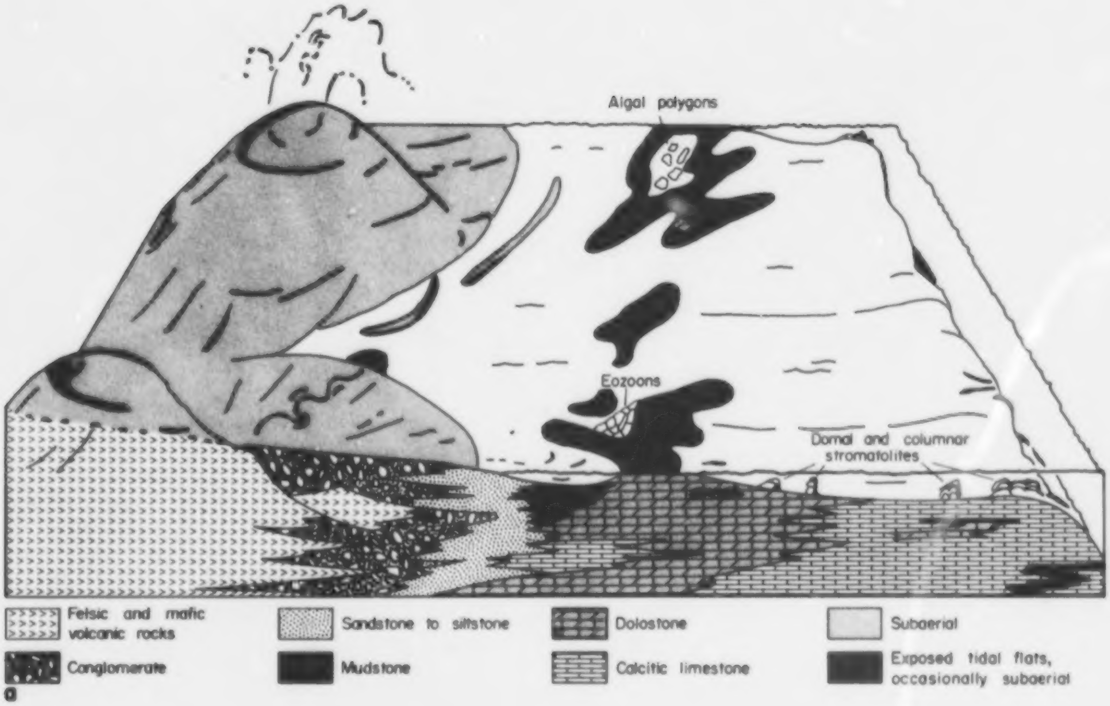

78°W
Carbonate shelves
Geological boundary
Domain boundary
Terrane boundary
Quebec
N
Ottawa River
Central Gneiss Belt
Central Metasedimentary Belt Boundary Zone
Bancroft Terrane
Arnprior
Bancroft
45°N
Minden
Elzevir Terrane
Mazinaw Terrane
Sharbot Lake Terrane
Perth
Madoc
Frontenac Terrane
0
50 km
78°W
77°W
78°W
Carbonate turbidite, turbidite fans, local shelves
Unsubdivided
Geological boundary
Domain boundary
Terrane boundary
Quebec
N
Ottawa River
Central Gneiss Belt
Central Metasedimentary Belt Boundary Zone
Bancroft Terrane
Arnprior
Bancroft
45°N
Minden
Elzevir Terrane
Mazinaw Terrane
Sharbot Lake Terrane
Perth
Madoc
Frontenac Terrane
0
50 km
78°W
77°W

**Figure 19.69.** Sedimentary assemblages in the Central Metasedimentary Belt.


Algal polygons
Eozoons
Domal and columnar stromatolites
Felsic and mafic volcanic rocks
Sandstone to siltstone
Dolostone
Subaerial
Conglomerate
Mudstone
Calcitic limestone
Exposed tidal flats, occasionally subaerial
a

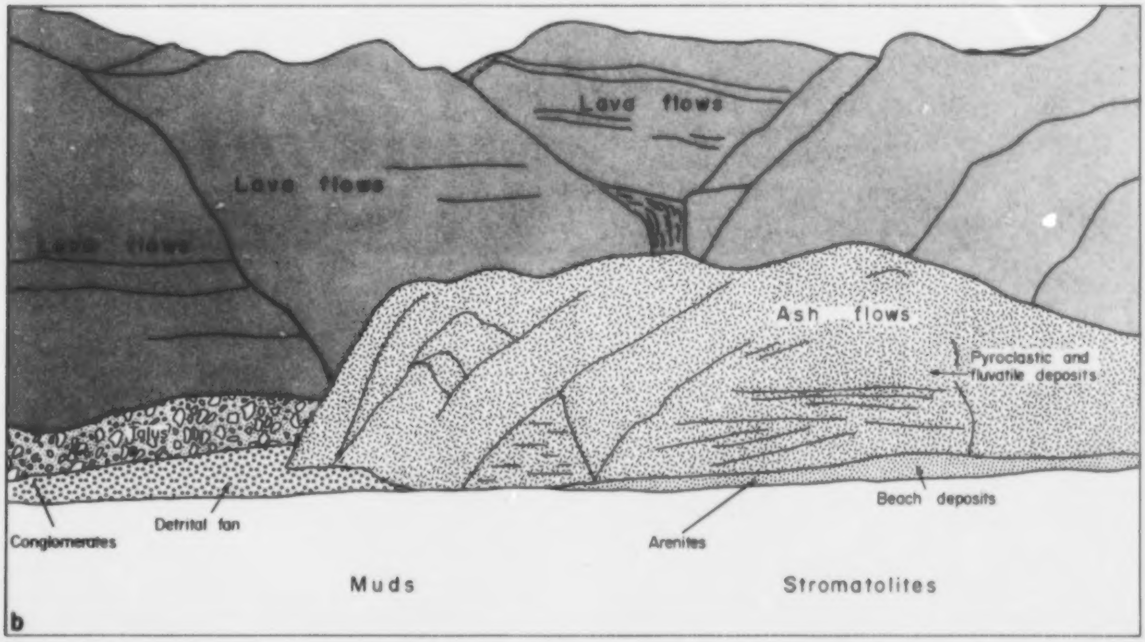

Lava flows
Lava flows
Lava flows
Ash flows
Pyroclastic and fluvatile deposits
Talus
Beach deposits
Conglomerates
Detrital fan
Arenites
Muds
Stromatolites
b

**Figure 19.70b.** Sketch showing possible depositional environments present at land-sea interface around volcanic islands and interrelationship between volcanic and sedimentary rocks. The Belmont Lake Formation (Belmont Domain) sedimentary rocks may have been deposited in this setting.

**Table 19.16.** Summary chart showing characteristic features of Central Metasedimentary Belt plutonic suites.

| Rock Suite | Age Range (Ma) | Rock Types |
|---|---|---|
| Fenite-carbonatite,<br>granite pegmatite | 1070–1040 | "fenite" or "skarn" assemblages, calcite apatite veins, carbonatite, pegmatitic granite, granite pegmatite<br>mineralization: U-Th, Mo, rare-element pegmatites, feldspar, mica |
| Monzonite-diorite<br>**Skootamatta Suite** | 1090–1075 | potassic monzonite and syenite, biotite diorite, minor granite, nepheline (ne)- normative monzodiorite, peridotite, lamprophyre |
| Syenite-monzonite<br>**Gananoque Suite** | 1180–1160 | potassic syenite and monzonite; minor syenogranite, monzogranite, biotite diorite, gabbro |
| Alaskite<br>**Methuen Suite** | 1250–1240 | alaskite; minor alkalic feldspar syenite albite granite, syenogranite, aplite<br>mineralization: contact metasomatic skarn deposits, high-fluorite granites may host Nb-Ta deposits |
| Diorite-gabbro<br>**Lavant Suite** | 1250–1240 | diorite, gabbro; minor quartz diorite, tonalite, granodiorite, albite syenite and granite, pyroxenite.<br>mineralization: Fe-Ti magmatic sulphide deposits, contact metasomatic iron deposits |
| Tonalite<br>**Elzevir Suite** | 1280–1270 | tonalite, granodiorite, trondhjemite; minor diorite, albite granite |
| Gabbro<br>**Killer Creek Suite** | 1300–1275 | gabbro, melanocratic gabbro, anorthositic gabbro, hornblendite, minor anorthosite<br>mineralization: Cu-Ni magmatic sulphide deposits, possible Cr and Ti deposits |
| Anorthosite | 1280–1250 | oligoclase to labradorite anorthosite; minor gabbro, diorite, tonalite, quartz syenite, syenite<br>mineralization: alumina sources |
| Nepheline syenite | 1280–1250 | nepheline syenite, urtite, malignite, ne-normative diorite and gabbro, ne-normative albite syenite; minor ijolite, melteigite, nepheline diorite, gabbro<br>mineralization: alumina source, magnetite concentrations |
| Early tonalite<br>**Dysart Suite** | 1370–1340 | tonalite, trondhjemite, granodiorite |

respectively, suggesting relatively slow cooling after the peak of the Ottawan Orogeny metamorphism at *circa* 1080 Ma. It is not known if a second regional metamorphic event occurred in the Bancroft Terrane in association with the *circa* 1070 million-year-old Ottawan Orogeny. The fenite-carbonatite-pegmatite event may have been partly responsible for prolonging the cooling period until after 1000 million years, and thus the 1045 to 1020 million-year-old metamorphism recorded by Mezger et al. (1990) may represent the termination of a longer metamorphic episode.

## Structure

The structural style of the Bancroft Terrane, particularly in the extreme northern and western parts of the terrane, is similar to that observed in the CMBBZ, characterized by strong, linear transposed lithologic units, abundant development of marble tectonic breccias and isoclinal and sheath folds, the presence of thin, sheet-like plutonic nappes (e.g., Redstone Lake and Dysart tonalite gneisses, Hanmer 1988 and the Percy Lake meta-anorthosite, Marmont 1991) and a prominent, southeast-plunging lineation (Hewitt 1962, Hanmer and Ciesielski 1984; Hanmer et al. 1985). With increasing distance from the CMBBZ, lithologic units become more varied in trend, plutons take on more ovoid shapes and folds become more upright, although a northerly and westerly vergence is still common. Detailed structural studies within the Bancroft Terrane have been conducted by Appleyard (1974b), Appleyard and Stott (1975), Culshaw (1983) and Breaks and Thivierge (1985). Most have focussed on the area north of Bancroft.

## Mineral Deposits

### STRATIFORM ZINC DEPOSITS

The largest zinc deposit in the Bancroft Terrane is the Cadieux (Renprior) zinc deposit located about 7 km south of Renfrew which was first discovered in 1922. The deposit contains roughly 750 000 t of mineralization grading about 10% Zn and 1% Pb (Easton et al. 1986). The deposit is hosted in dolomitic and calcitic marbles bounded to the northwest by siliceous clastic metasedimentary rocks and by granite and diorite to the southeast. Sangster (1991) reports $^{34}$S values of -2.5‰ for the Cadieux deposit and suggests an origin involving exhaled basinal brines interacting with carbonate rocks with sulphur being derived from adjacent metavolcanic rocks. This deposit occurs near the poorly defined boundary between the Mazinaw and Bancroft terranes. A metavolcanic component to the sulphur isotope data (Sangster 1991) is consistent with the general increase in interbedded volcanic and volcaniclastic material that occurs in the eastern and southern Bancroft Terrane.

The Salerno Lake zinc deposit near Furnace Falls is smaller. Grade and tonnage for the deposit are unavailable; assessment file reports indicate grades of 3% Zn (*see* Easton 1987a). The deposit is hosted in interbedded dolomitic (commonly tremolitic) and calcitic marbles near the eastern margin of the Salerno Lake Gabbro. Easton (1987a) reports a zone of arsenic enrichment in the vicinity of the deposit. High $^{34}$S values (greater than 10‰) for the Salerno Lake deposit (Sangster 1991) are known from only a few zinc deposits in the Grenville, most notably Balmat–Edwards. The stratiform zinc deposits in the Bancroft Terrane are directly comparable to the Balmat–Edwards deposit (in New York State) in stratigraphic setting, rock types and sulphur isotope data.

Other zinc occurrences in the Bancroft Terrane are also associated with dolomitic marbles, commonly tremolitic, locally stromatolitic and include the Buller Lake area in the Denna Lake Structural Complex (Easton 1987a, 1990a) and the Silver Mountain occurrence near Furnace Falls (Easton 1987a). Deposits and occurrences to date are mainly reported from the better mapped southern part of the terrane, particularly where it is least disrupted tectonically.

### METASOMATIC IRON DEPOSITS

Iron and iron-titanium iron skarn deposits are common in Haliburton Group marbles within 1 to 2 km of mainly gabbroic, but also some diorite and nepheline-syenite intrusions. Several in the Furnace Falls area were mined for iron in the late 1800s, producing between 500 and 2000 t each, including the Howland, Imperial, Paxton and Victoria mines (Rose 1958; Easton 1987a, 1987e). The Paxton Mine

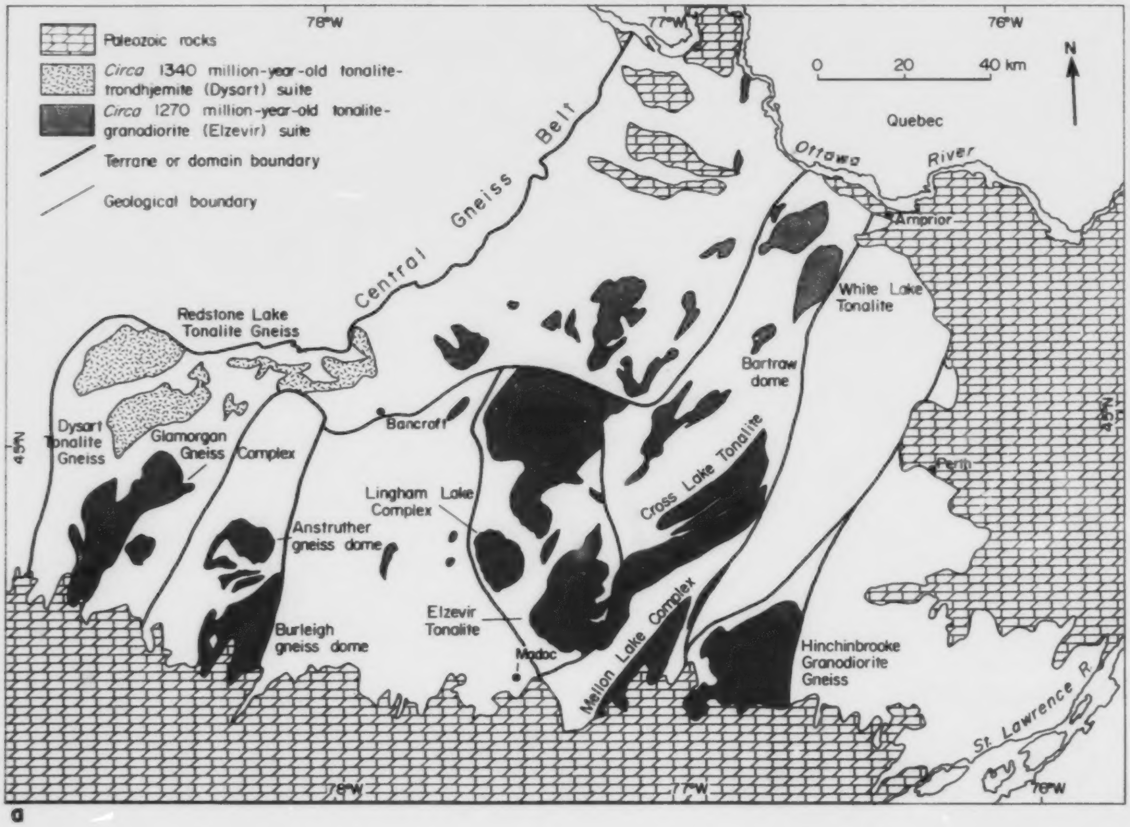

Paleozoic rocks
*Circa* 1340 million-year-old tonalite-trondhjemite (Dysart) suite
*Circa* 1270 million-year-old tonalite-granodiorite (Elzevir) suite
Terrane or domain boundary
Geological boundary
78°W
77°W
76°W
0
20
40 km
N
Quebec
Ottawa River
Arnprior
Central Gneiss Belt
White Lake Tonalite
Redstone Lake Tonalite Gneiss
Bartraw dome
Bancroft
Dysart Tonalite Gneiss
Glamorgan Gneiss Complex
45°N
Cross Lake Tonalite
Perth
Lingham Lake Complex
Anstruther gneiss dome
Elzevir Tonalite
Burleigh gneiss dome
Madoc
Mellon Lake Complex
Hinchinbrooke Granodiorite Gneiss
St. Lawrence R.
a

**Figure 19.71a.** Distribution of plutonic suites (*modified from* Lumbers et al. 1990) within the Central Metasedimentary Belt. For description of terrane and domain boundaries *see* Figure 19.51a. Distribution of early Dysart Suite (1340 Ma) and late (1270 Ma) Elzevir Suite plutonic rocks.

is located within the Denna Lake Structural Complex. The Victoria Mine is unusual in that it contains 1 to 2 wt % light rare earths (Easton 1987a).

## MAGMATIC IRON-TITANIUM AND COPPER-NICKEL DEPOSITS

These deposits are associated with gabbroic rocks of the diorite-gabbro and nepheline syenite suite and syenites of the nepheline syenite suite. Mineralization in the iron-titanium deposits is generally magnetite, with titanite containing most of the 1 to 3% $TiO_2$ content reported in these occurrences (Carter 1984). Mineralization occurs either as scattered pods of massive and disseminated magnetite (Boulter, Fraser Lake and Bow Lake deposits), as parallel layers and lenses of disseminated to massive magnetite (Grattan deposit), or large zones of disseminated magnetite within syenitic and nephelinitic gneisses (Coltson Lake, Trent River, White Foot Lake and Bodies "B", "C" and "D" *of* Easton 1987a). The Grattan is the largest deposit, containing 9 million t of 27.74% Fe, 0.78 to 1.84 wt % $TiO_2$ and 0.14 to 1.30 wt % $P_2O_5$ (Carter 1984).

The other iron-titanium deposits are hosted in gabbroic intrusions and consist of magnetite, ilmenite and ilmeno-magnetite as disseminated to massive layers and lenses, parallel to the contacts of large intrusions and relict igneous layering in the intrusions (Carter 1984). Deposits include the Keller and Pusey deposits (Carter 1984).

The Simmons deposit (Carter 1984) is the largest gabbroic-hosted copper-nickel deposit within the Bancroft Terrane and is hosted in the Raglan Hills gabbro. It is associated with a roof pendant of possible metavolcanic rocks. Magnetite-bearing amphibolites within the gabbro locally host gold mineralization, and comparisons have been made with the New Calumet deposit near the Ontario–Quebec boundary. Other gabbroic hosted copper-nickel deposits in the Bancroft Terrane include the Ameranium, Genricks Lake, Landolac and Raglan deposits within the Raglan Hills gabbro. Mineralization in all consists of pyrrhotite, chalcopyrite, pentlandite and pyrite (Carter 1984).

## GOLD DEPOSITS

No gold deposits are reported from the Bancroft Terrane by Carter (1984) or Sangster and Bourne (1982). A few minor occurrences in quartz veins are known locally from the terrane but are not extensive (Easton 1987a).

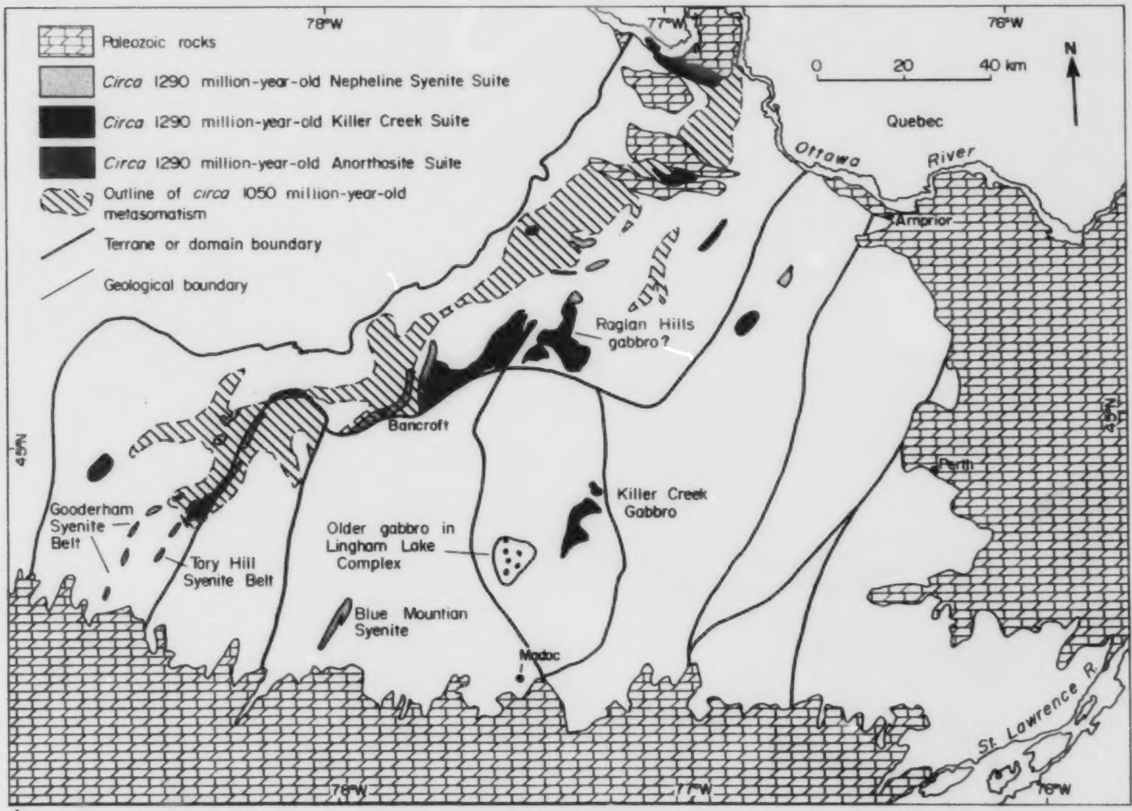

Paleozoic rocks
*Circa* 1290 million-year-old Nepheline Syenite Suite
*Circa* 1290 million-year-old Killer Creek Suite
*Circa* 1290 million-year-old Anorthosite Suite
Outline of *circa* 1050 million-year-old metasomatism
Terrane or domain boundary
Geological boundary
78°W
77°W
76°W
45°N
0
20
40 km
N
Quebec
Ottawa River
Arnprior
Raglan Hills gabbro?
Bancroft
Perth
Killer Creek Gabbro
Gooderham Syenite Belt
Tory Hill Syenite Belt
Older gabbro in Lingham Lake Complex
Blue Mountian Syenite
Madoc
St. Lawrence R.
b

**Figure 19.71b.** Distribution of plutonic suites (*modified from* Lumbers et al. 1990) within the Central Metasedimentary Belt. For description of terrane and domain boundaries *see* Figure 19.51a. Distribution of anorthosite suite and Killer Creek Suite (ca. 1290 to 1250 Ma) and nepheline syenite suite (ca. 1290 to 1250 Ma) plutonic rocks.

## SULPHIDE DEPOSITS

Pyrite, with minor pyrrhotite, chalcopyrite, arsenopyrite and pentlandite occurrences are found locally in the terrane, commonly as contact metasomatic concentrations in marbles adjacent to diorite plutons of the Lavant Suite. Locally, these deposits have been concentrated due to pre-Paleozoic weathering, for example, the Hughes deposit near Furnace Falls (Easton 1987a). These deposits are small, and significant pyrite production has only been reported from sulphide deposits within the Elzevir and Mazinaw terranes.

## MOLYBDENUM DEPOSITS

Over 100 molybdenum-bearing metasomatic deposits are reported from the Bancroft Terrane (Eardley-Wilmont 1925; Carter 1984). Most deposits are of high grade, but low tonnage. Three main types of occurrences are reported (Karvinen 1973; Carter 1984; Lentz 1991b), namely: 1) strata-bound deposits hosted in pyroxene-rich skarns; 2) deposits hosted in granitic pegmatite dikes or sills; and 3) stratiform deposits associated with amphibolitic gneisses. In all types, granitic pegmatites are spatially associated with the deposits, as are uranium-bearing pegmatites and deposits. Easton (1987a, 1990a) outlined areas within the Denna Lake Structural Complex with greater molybdenum potential, noting a dolomite-amphibolite-granite pegmatite association in molybdenum-rich areas. Lentz (1991a, 1991b) interrelates all 3 deposit types, with mineralization ascribed to the post-Grenvillian alkali metasomatic, fenite-carbonatite-pegmatite event.

## URANIUM AND RARE EARTH MINERALIZATION

Uranium-, molybdenum- and REE-bearing pegmatites, skarns and fluorite-apatite-calcite veins occur throughout the Central Metasedimentary Belt but are most abundant in the Bancroft and northern Elzevir terranes. These deposits are subeconomic, but in the past they have been mined for apatite (1860–1890), molybdenum (1914–1920), phlogopite (1890–1960), muscovite and feldspar (1890–1970), fluorite (1940–1950) and most recently uranium (mid-1950s and 1970 to early 1980). The economic viability of these deposits declined because of changing demand, low tonnages, low grade and erratic distribution of mineralization. In addition to the traditional metals associated with these deposits (uranium and molybdenum), thorium, yttrium, beryllium, rare-earth elements (REE) and niobium (Ellsworth 1932; Lang 1952; Lang et al. 1962; Shaw 1958; Rose 1960; Hewitt 1967a, 1967b, 1967c; Dawson 1974)

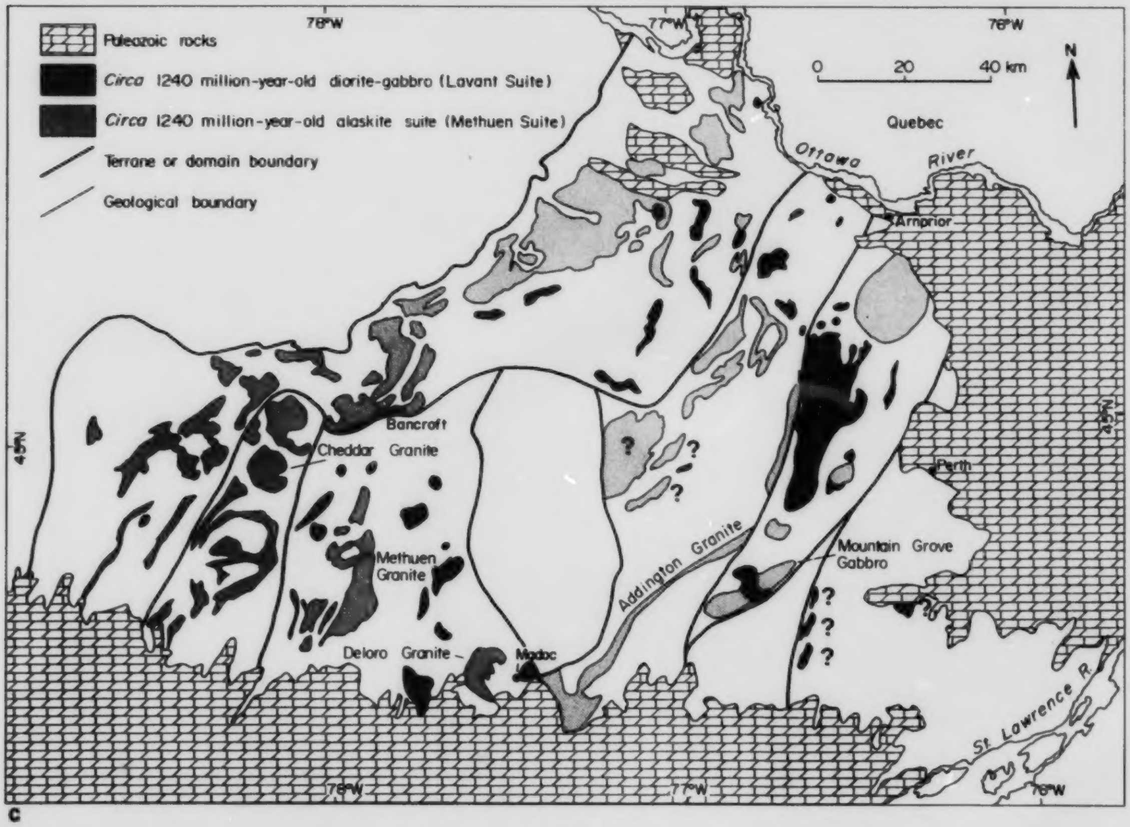

Paleozoic rocks
Circa 1240 million-year-old diorite-gabbro (Lavant Suite)
Circa 1240 million-year-old alaskite suite (Methuen Suite)
Terrane or domain boundary
Geological boundary
78°W
77°W
76°W
0
20
40 km
N
Quebec
Ottawa River
Arnprior
Perth
45°N
45°N
Bancroft
Cheddar Granite
Methuen Granite
Deloro Granite
Madoc
Addington Granite
Mountain Grove Gabbro
St. Lawrence R.
?
?
?
?
?
?
78°W
77°W
76°W
c

**Figure 19.71c.** Distribution of plutonic suites (*modified from* Lumbers et al. 1990) within the Central Metasedimentary Belt. For description of terrane and domain boundaries *see* Figure 19.51a. Distribution of Lavant Suite diorite-gabbro (ca. 1240 Ma) and Methuen Suite alaskite (ca. 1240 Ma) plutonic rocks.

have anomalous concentrations in the pegmatites, skarns and fluorite-apatite-calcite veins.

Lentz (1991b) proposes that, in general, each of the uranium, molybdenum and rare-element deposit types are spatially and temporally related based on geologic, textural, mineralogic, geochronologic and geochemical data (Satterly 1957; Shaw 1958; Hewitt 1967b; Masson and Gordon 1981). There is no consensus, however, on the exact relationship between the pegmatites, skarns and fluorite-apatite-calcite veins.

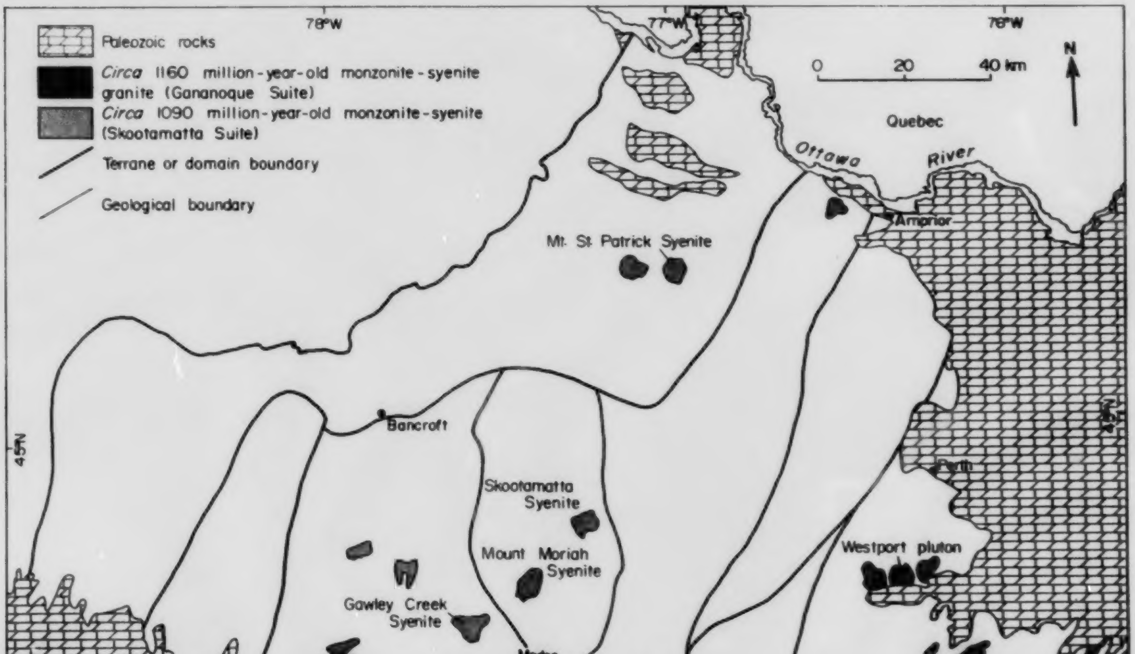
## BUILDING STONE

Several stone prospects have been identified in the southwestern Bancroft Terrane (Zeeman and Easton 1985; Easton 1990a; Marmont 1991) and include units in the Glamorgan Gneiss Complex and some of the more massive marble bands. Stone prospects in the Bancroft area are summarized *in* LeBaron et al (1990) and Hewitt (1964c) and in the Renfrew area by Storey and Vos (1981a). In addition, several flagstone quarries are present along the western boundary of the Bancroft Terrane. Hewitt (1964d), Easton (1990a, 1987e) and Marmont (1991) have described these occurrences.

Marble has been quarried from several sites in the Bancroft Terrane (Goudge 1938; Hewitt 1964c; Easton 1987a, 1987e), the largest of which are the dolomite quarries near Eagle Lake (near Minden) and Haley Station (near Renfrew). In the Renfrew area, the dolomite marble is used in the production of magnesium metal (Easton et al. 1986).

## OTHER INDUSTRIAL MINERALS

The Black Donald graphite mine operated between 1896 and 1954, producing 20.45 million kilograms of coarse flake graphite before it was flooded during the creation of Centennial Lake in 1954 (Spence 1920; Hewitt 1965; Storey and Vos 1981b). The graphite zone is 1.5 to 1.8 m thick and is conformable with layering in the host calcitic and dolomitic marbles and calc-silicate metasedimentary rocks (Spence 1920). The origin of the zone is unknown. Several smaller graphite occurrences are present in the vicinity of the Black Donald Mine (Lumbers 1982). Easton (1990a) described several small graphite occurrences in the Denna Lake Structural Complex. Deformation zones associated with dolomitic marbles within the terrane are potential



78°W
77°W
76°W
▨ Paleozoic rocks
■ *Circa* 1160 million-year-old monzonite-syenite granite (Gananoque Suite)
▩ *Circa* 1090 million-year-old monzonite-syenite (Skootamatta Suite)
Terrane or domain boundary
Geological boundary
0 20 40 km
N
Quebec
Ottawa River
Arnprior
Mt. St. Patrick Syenite
Bancroft
Skootamatta Syenite
Mount Moriah Syenite
Gawley Creek Syenite
Madoc
Perth
Westport pluton
45°N
78°W
77°W
76°W
d

**Figure 19.71d.** Distribution of plutonic suites (*modified from* Lumbers et al. 1990) within the Central Metasedimentary Belt. For description of terrane and domain boundaries *see* Figure 19.51a. Distribution of Gananoque Suite monzonite-syenite-granite (ca. 1160 Ma) and Skootamatta Suite (1090 Ma) syenite plutonic rocks.

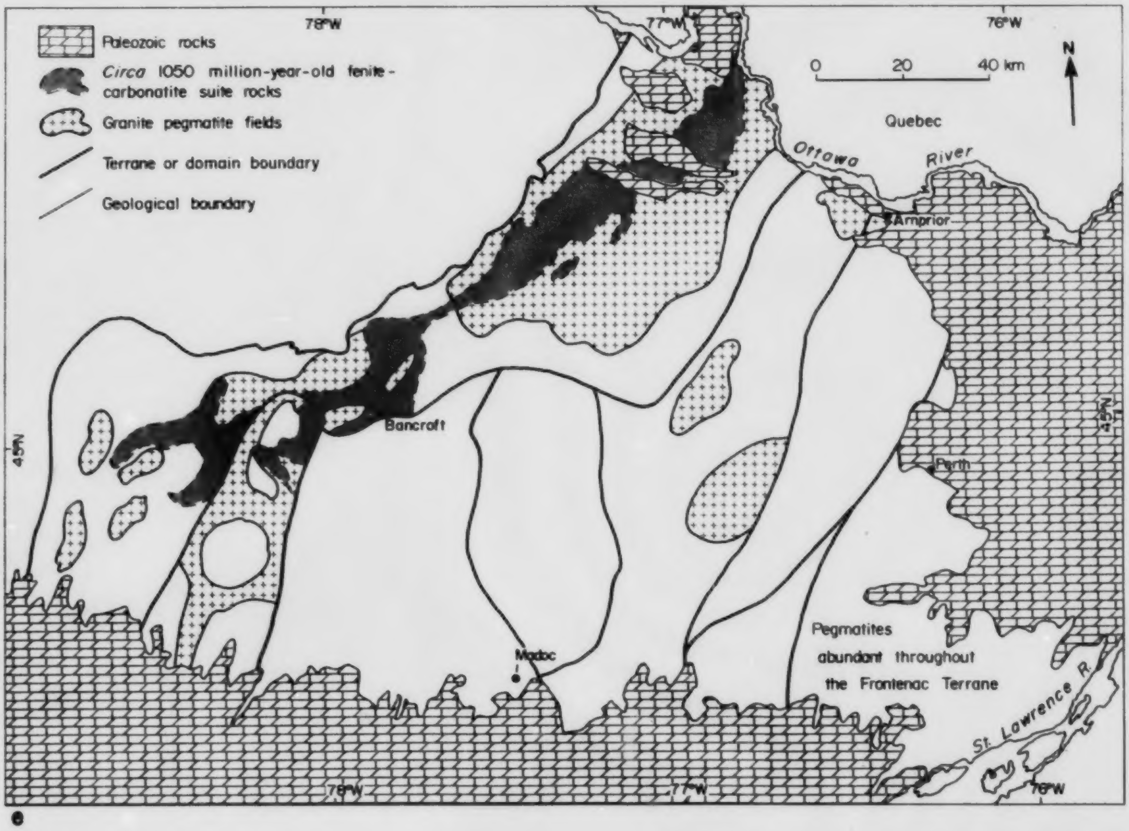
exploration targets for graphite. Lumbers (1982) drew attention to the association between rusty schists and graphite mineralization in the Llano Uplift in Texas and suggested that rusty schists in contact zones of granitic plutons within the Bancroft Terrane may be potential targets for graphite exploration.

The Ruby Mine in Ashby Township produced 360 tons of andradite garnet (Eardley-Wilmont 1927; Atchison 1937; Satterly 1945; Evans 1964). Corundum occurs in several parts of the terrane, in association with the nepheline syenite suite. Past production has come from the Craigmont Mine in Raglan Township (Satterly 1945). Lumbers (1982) notes that the small size and scattered nature of the deposits in the Renfrew area has prevented profitable mining of the deposits in the past. Satterly (1945) and Hewitt (1961) discuss nepheline syenite occurrences within the terrane, however none are of sufficient size or purity to be of economic interest. Dolan et al. (1991) summarizes potential alumina sources from nepheline syenite and anorthositic host rocks in the Grenville Province; most of these occurrences are located in the Bancroft Terrane.

Feldspar and mica were locally exploited in small quantities from pegmatite veins throughout the terrane near the turn of the century. Many of these pegmatites, as well as the calcite-vein dikes of the Bancroft region, host minerals of interest to collectors, and many are summarized in a series of mineral collecting guidebooks produced by the Geological Survey of Canada (Sabina 1983, 1986a, 1986b, 1987). Masson and Gordon (1981) summarize the radioactive-bearing pegmatite occurrences, and Storey and Vos (1981b) summarize the industrial mineral prospects of Renfrew County.

## ELZEVIR TERRANE

The Elzevir Terrane can be divided into 3 lithotectonic domains, each characterized by distinctive stratigraphy and structural style (*see* Figures 19.51 and 19.52, and Table 19.5). However, on a broader scale, these domains are sufficiently similar in stratigraphy, plutonic history and geologic history, that they are best regarded as subsets of a larger terrane. The domains are outlined in Figure 19.51a and can be regarded as parts of the Elzevir Terrane. Further, the Sharbot Lake Terrane is similar to the Elzevir Terrane in terms of rock types and history. Domain boundaries are tectonic and may be marked by straight gneiss zones, shear zones or late extensional mylonite zones (*see* Figure 19.51a). In a broad sense, the Elzevir Terrane is



e

**Figure 19.71e.** Distribution of plutonic suites (*modified from* Lumbers et al. 1990) within the Central Metasedimentary Belt. For description of terrane and domain boundaries *see* Figure 19.51a. Distribution of carbonatite-fenite (1070 to 1040 Ma) and granite pegmatite suite (1060 to 1030 Ma) rocks. Major pegmatite fields are also identified.

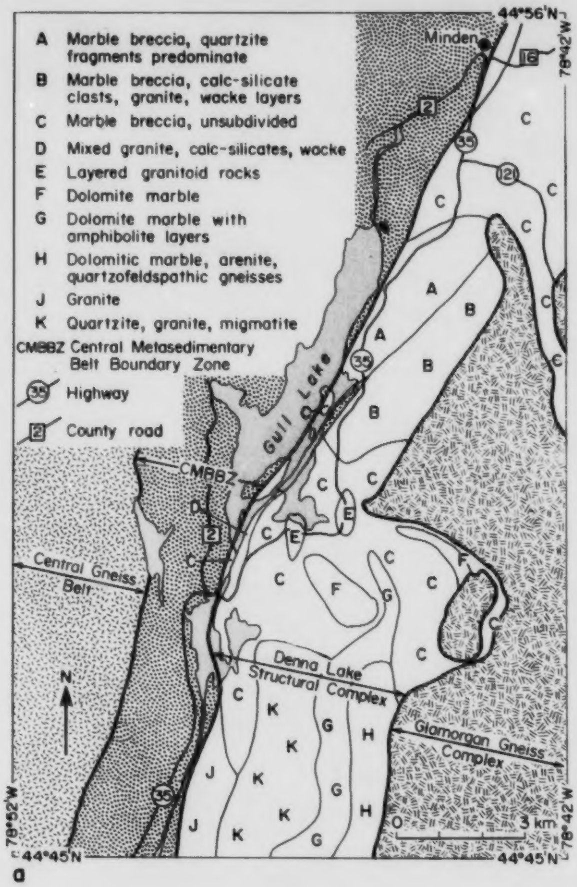

A Marble breccia, quartzite fragments predominate
B Marble breccia, calc-silicate clasts, granite, wacke layers
C Marble breccia, unsubdivided
D Mixed granite, calc-silicates, wacke
E Layered granitoid rocks
F Dolomite marble
G Dolomite marble with amphibolite layers
H Dolomitic marble, arenite, quartzofeldspathic gneisses
J Granite
K Quartzite, granite, migmatite
CMBBZ Central Metasedimentary Belt Boundary Zone
35 Highway
2 County road
Minden
44°56'N
78°42'W
Gull Lake
CMBBZ
Central Gneiss Belt
Denna Lake Structural Complex
Glamorgan Gneiss Complex
N
0 3 km
78°52'W
44°45'N
a

**Figure 19.72a.** "Ghost stratigraphy" of the Haliburton Group preserved within tectonically disrupted strata of the Denna Lake Structural Complex in the Minden area (*modified from* Easton 1990a).

lithologically, structurally and metamorphically simple (monocyclic) compared to more complex and probably polycyclic terranes such as the Mazinaw and Frontenac terranes. Table 19.7 and Figures 19.53 and 19.54 summarize the stratigraphic sequences in the different domains and highlight the lithologic differences between the domains.

## Harvey–Cardiff Arch

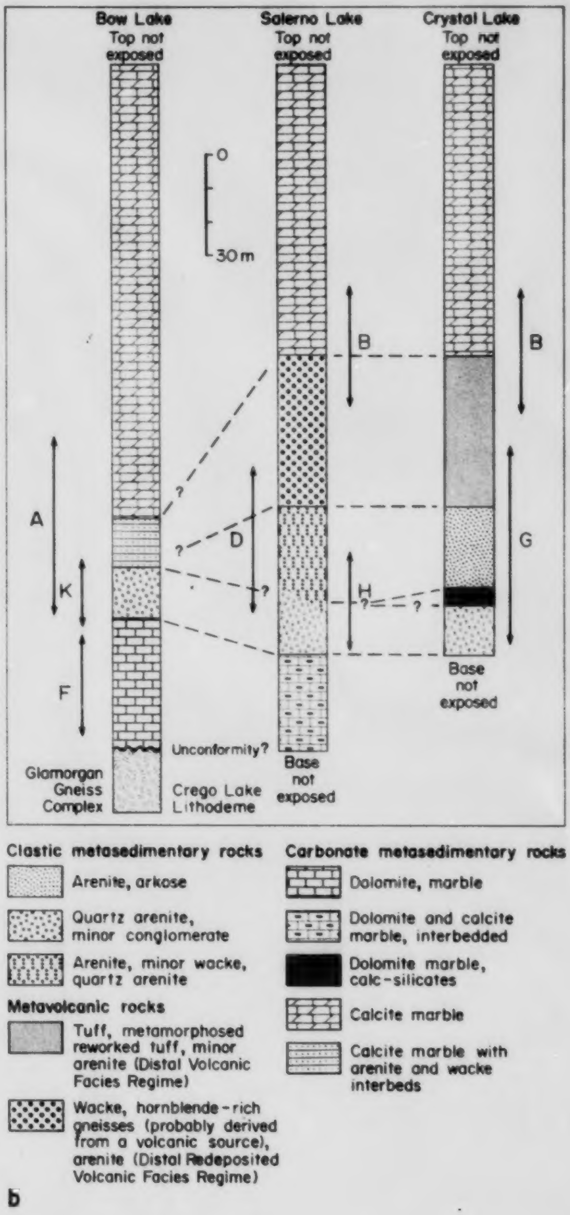

Bow Lake
Top not exposed
Salerno Lake
Top not exposed
Crystal Lake
Top not exposed
0
30 m
A
K
F
B
D
H
G
?
Unconformity?
Glamorgan Gneiss Complex
Crego Lake Lithodeme
Base not exposed
Base not exposed
Clastic metasedimentary rocks
Arenite, arkose
Quartz arenite, minor conglomerate
Arenite, minor wacke, quartz arenite
Metavolcanic rocks
Tuff, metamorphosed reworked tuff, minor arenite (Distal Volcanic Facies Regime)
Wacke, hornblende-rich gneisses (probably derived from a volcanic source), arenite (Distal Redeposited Volcanic Facies Regime)
Carbonate metasedimentary rocks
Dolomite, marble
Dolomite and calcite marble, interbedded
Dolomite marble, calc-silicates
Calcite marble
Calcite marble with arenite and wacke interbeds
b

**Figure 19.72b.** Comparison of stratigraphic sections from Howland area (undisrupted Grenville Supergroup strata) with dominant metasedimentary sequences present within the Denna Lake Structural Complex. Letters A to K refer to zones shown in (a). Note "ghost stratigraphy" present in Denna Lake Structural Complex.

The Harvey–Cardiff Arch is characterized by domal plutonic masses ringed by metavolcanic and metasedimentary rocks. The arch has been recently mapped in detail by several workers, including Morton (1983) and Bright (1976, 1977, 1987a, 1987b, 1988) and has also been the subject of several structural and geochemical studies (e.g., Fowler and Doig 1979, 1983; Culshaw 1983). It is still unclear if the gneiss domes of the arch are classic mantled gneiss domes, or do they simply represent the exposed roof of a large, possibly diapiric, plutonic mass. The contact between the Bancroft Terrane and the Harvey–Cardiff Arch is approximated by the Bancroft shear zone (Carlson et al. 1990). In the east, the Burleigh fault and the McArthurs Mills shear zone separates the Harvey–Cardiff Arch from the Belmont Domain.

The Harvey–Cardiff Arch differs from the Bancroft Terrane in several respects: 1) it contains more volcanic rocks; 2) metasedimentary rocks in the arch are less mature, being dominated by metawackes and turbiditic metacarbonates; 3) the structural style is different. The prominent southeast-plunging lineation typical of the

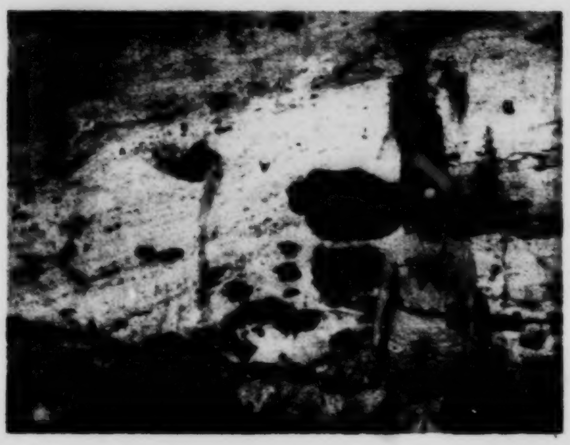


**Figure 19.73.** Marble tectonic breccia, Denna Lake Structural Complex, Highway 121 south of Minden. Breccia at this locality is heterogeneous and contains a wide variety of fragments.

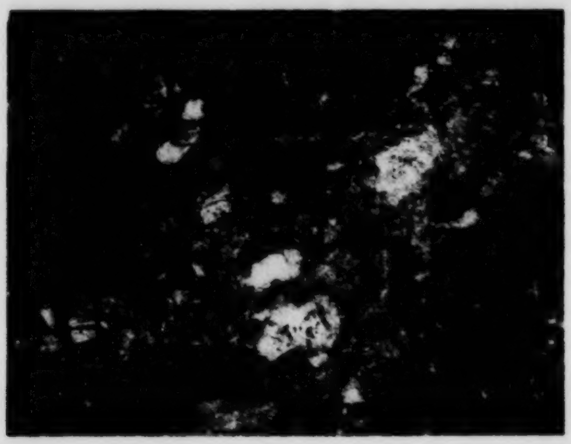


**Figure 19.74.** Marble tectonic breccia, Denna Lake Structural Complex, Highway 35 south of Minden. Breccia at this locality is monomictic and contains only quartz arenite fragments. Hammer handle is 30 cm long.

Bancroft Terrane is less well-developed, and tonalitic rocks have domal, rather than elongated, nappe-like forms; and 4) nepheline syenite and carbonatite-fenite suite rocks are apparently absent from the arch. Despite these differences, the Harvey–Cardiff Arch may be related to the Bancroft Terrane. The changes in the supracrustal rock types could be viewed as a change from the shelf environment of the Bancroft Terrane to a somewhat deeper water, volcanic and tectonically active basin. (Note: this does not necessarily imply that the Bancroft Terrane and the Harvey–Cardiff Arch were con-tiguous.) The change in structural style may simply reflect the fact that the 2 regions had somewhat different tectonic histories. The fact that both areas contain significant pegmatite-hosted uranium mineralization also suggests that the Bancroft Terrane and the Harvey–Cardiff Arch were linked during the later stages of their tectonic evolution.

## STRATIGRAPHIC RELATIONSHIPS

Bright (1976, 1977) proposed several new formations in the Burleigh and Eels Lake areas, as summarized in Table 19.10. Many of Bright's stratigraphic units can be traced into the Harvey Township area mapped by Morton (1983). The stratigraphic unit of greatest significance introduced by Bright (1976, 1977) is the Anstruther Lake Group, which consists of leucocratic quartzofeldspathic gneisses of roughly granitic composition, which rim the migmatitic tonalitic cores of the gneiss domes in the Harvey–Cardiff Arch. Bright (1976, 1977) interpreted the Anstruther Lake Group as a metasedimentary sequence developed unconformably on a tonalitic gneissic basement now preserved as the cores of the gneiss domes. Carter (1984) loosely correlated this sequence with the "basal coarse clastic" sequence of Lumbers (1982) present along the Bancroft-CMBBZ boundary in the Renfrew area. Easton (1987a, 1987e, 1990a) disagreed with this interpretation and concluded that the gneisses were probably metaigneous rocks, locally subjected to high strain. Further, the lack of a nearby source for this thick arkosic sequence makes a metasedimentary origin for these gneisses unlikely. Given the uncertain origin for the unit, it is proposed, here, that the term Anstruther Lake Group be replaced by the term Anstruther Lake Lithodeme, thereby removing any genetic connotation and retaining the term as a useful map unit.

Other stratigraphic units proposed by Bright (1976, 1977) are outlined in Table 19.10. Bright (1976, 1977) observed that the distribution of uraniferous mineralization in late pegmatite veins in the region is stratigraphically controlled, with all significant uranium deposits being hosted in metavolcanic rocks or volcaniclastic metasedimentary rocks of the Eels Lake Formation.

## GEOCHEMISTRY OF THE METAVOLCANIC ROCKS

Morton (1983) reports chemical data from a diverse sequence of volcanic rocks from Harvey Township on the western margin of the Harvey–Cardiff Arch. Basaltic rocks in the area are mainly tholeiites, locally nepheline normative, a characteristic Morton (1983) attributed to sodium enrichment during secondary alteration. Feldsparphyric basalts in the area have slightly ultramafic compositions and are probably sills or dikes (Morton 1983). Felsic volcanic rocks are dacitic to rhyodacitic in composition and are tholeiitic. The succession is similar to the typical subalkalic bimodal basalt-dacite-rhyolite sequence prevalent throughout much of the Elzevir Terrane, particularly in the Belmont Domain. Bright (1987a) reports similar results from the Cavendish area north of that studied by Morton (1983).

The eastern side of the Harvey–Cardiff Arch is characterized by tholeiitic mafic volcanic rocks (Bright 1987a, 1988). Dacitic and rhyodacitic volcanic rocks may be calc-alkalic, although too few samples were analyzed to characterize the chemical variation.

## PLUTONISM

Two pluton types are present in the area, both forming domal, tonalitic to granodioritic, commonly migmatitic, plutonic masses. They are: 1) the *circa* 1270 million-year-old Elzevir Suite, including the cores of the Burleigh gneiss dome and the Anstruther gneiss dome; and 2) alaskitic granites of the *circa* 1245 million-year-old Methuen Suite (e.g., Cheddar and Cardiff granites and granitic rocks rimming the Burleigh and Anstruther gneiss domes). Pegmatites of the *circa* 1050 million-year-old fenite-carbonatite-pegmatite Suite are also abundant. Fenites and carbonatites are poorly represented in the Harvey-Cardiff Arch in contrast to the Bancroft Terrane. A Rb-Sr age of 1207±46 Ma from the Cheddar Granite (Fowler and Doig 1983) is in rough agreement with zircon ages obtained from Methuen Suite granites from the Belmont Domain. Wu (1984) and Wu and Kerrich (1986) present chemical and oxygen isotopic data from the Cheddar Granite. Wu (1984) observed that the rare earth element patterns of the Cheddar Granite differed from normal granitic fractionation trends and suggested that the pluton was autometasomatized by a volatile heavy rare earth element-rich fluid. Oxygen isotope values from the pluton were uniform (8.8±0.9‰, Wu and Kerrich 1986), suggesting either complete homogenization during the metasomatizing event, or that the metasomatism did not affect primary oxygen isotope values. As the Cheddar Granite is located within the zone of *circa* 1050 million-year-old metasomatism documented by Lumbers et al. (1990), the suggestion that the oxygen data were homogenized during metasomatism seems plausible, and this later event may also be the cause of the unusual REE signature of the Cheddar Granite.

## STRUCTURAL HISTORY

The Harvey–Cardiff Arch is dominated by a series of orthogneiss-cored domes, surrounded by narrow belts of metavolcanic and metasedimentary rocks.

## METAMORPHIC HISTORY

The metamorphic history of the Harvey–Cardiff Arch has not been studied in detail. The arch is preserved at middle to upper amphibolite-facies conditions. No increase in metamorphic grade has been noted adjacent to the tonalite-cored domes present in the southern Harvey–Cardiff Arch, although the tonalite domes are locally migmatitic. The overall metamorphic history preserved in the arch does not

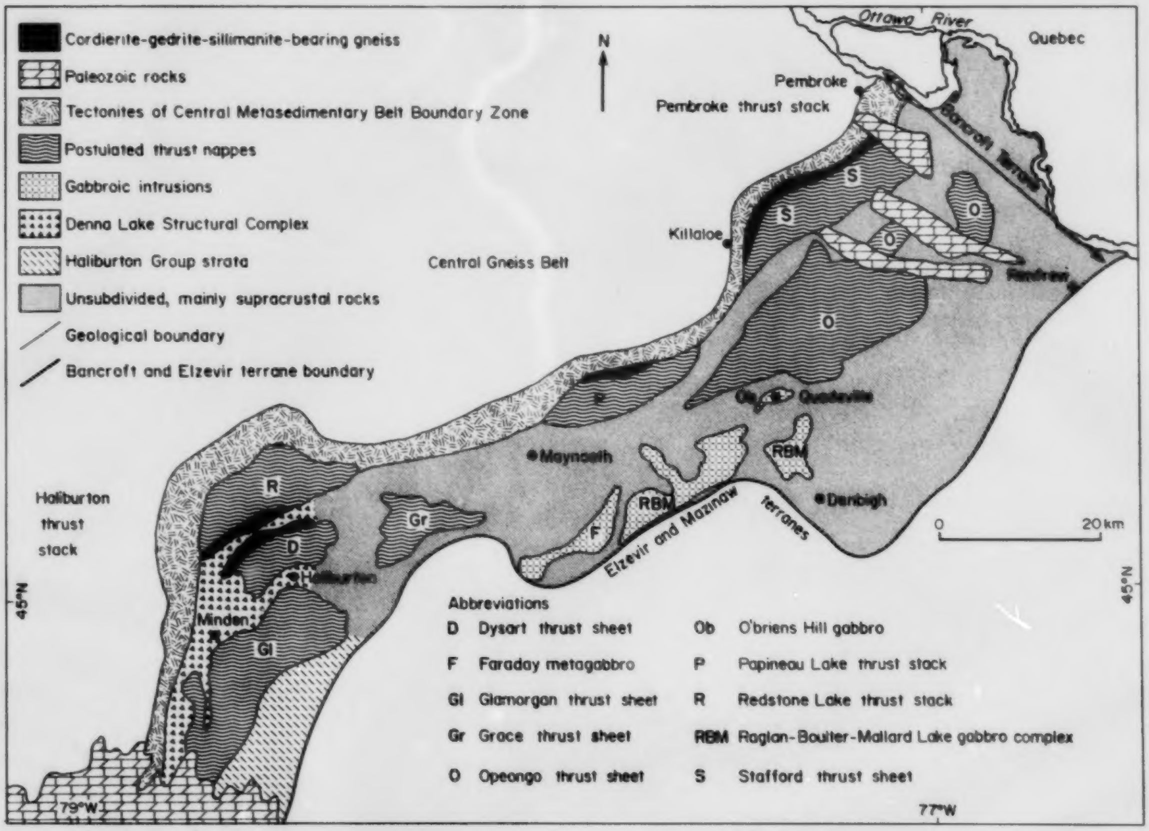

Cordierite-gedrite-sillimanite-bearing gneiss
Paleozoic rocks
Tectonites of Central Metasedimentary Belt Boundary Zone
Postulated thrust nappes
Gabbroic intrusions
Denna Lake Structural Complex
Haliburton Group strata
Unsubdivided, mainly supracrustal rocks
Geological boundary
Bancroft and Elzevir terrane boundary
N
Ottawa River
Quebec
Pembroke
Pembroke thrust stack
Bancroft Terrane
Killaloe
Renfrew
Central Gneiss Belt
S
S
O
O
O
P
Ob
Quadeville
RBM
Maynooth
RBM
Denbigh
F
Elzevir and Mazinaw terranes
R
Gr
D
Haliburton
Haliburton thrust stack
Minden
Gl
0
20 km
45°N
45°N
79°W
77°W
Abbreviations
D Dysart thrust sheet
F Faraday metagabbro
Gl Glamorgan thrust sheet
Gr Grace thrust sheet
O Opeongo thrust sheet
Ob O'briens Hill gabbro
P Papineau Lake thrust stack
R Redstone Lake thrust stack
RBM Raglan-Boulter-Mallard Lake gabbro complex
S Stafford thrust sheet

**Figure 19.75.** Distribution of cordierite-gedrite±sillimanite-bearing gneisses near the Bancroft Terrane–Central Metasedimentary Belt Boundary Zone boundary, in Ontario, and location of large thrust sheets (*modified from* McEachern 1990). Thickness of cordierite-gedrite±sillimanite gneiss zones is exaggerated. Distribution of possible thrust sheets of gabbro and gabbroic anorthosite along the southern boundary of the Bancroft Terrane is also shown. Thrust sheet names *from* Hanmer (1988) and McEachern (1990). Location of Haliburton Group supracrustals and the Denna Lake Structural Complex in the Minden–Haliburton area is shown for reference.

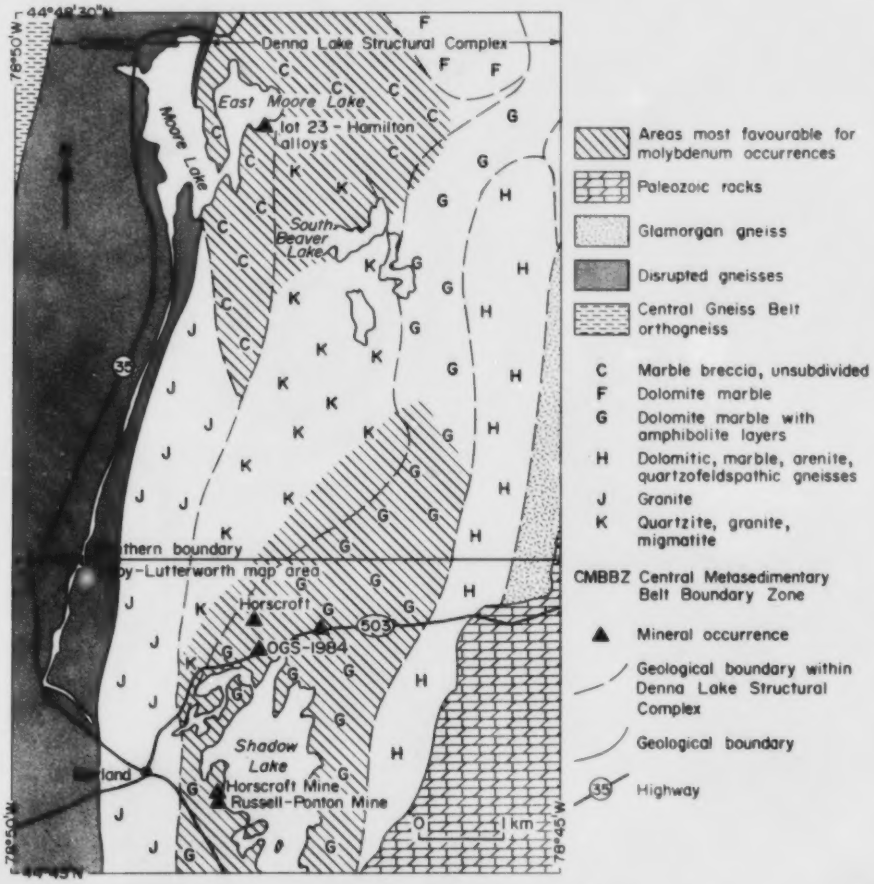
appear to differ from that preserved elsewhere in the Elzevir Terrane (Belmont Domain). At present, it appears that the Harvey–Cardiff Arch was subjected to a single intense regional metamorphic and deformational event, probably coincident with the Elzevirian Orogeny. The Ottawan Orogeny is reflected in resetting of K-Ar ages in the area.

## MINERAL DEPOSITS

Many mineral deposit types are absent from the Harvey–Cardiff Arch reflecting the absence of key rock types. The absence of gabbroic and dioritic intrusions means that copper-nickel and iron-titanium magmatic and contact metasomatic iron deposits are rare or absent in the arch. Bright (1977, 1987a, 1988) reports the presence of magnetite-pyrite-pyrrhotite mineralization in metasedimentary rocks of the Monmouth Formation and stratiform disseminated pyrite-pyrrhotite mineralization in metavolcanic rocks of the Eels Lake Formation. These deposits are generally small and scarce.

The most significant deposit type in the area are uranium-thorium deposits associated with late pegmatite veins, as well as molybdenum also found in pegmatite veins (Satterly 1957; Bright 1976, 1977, 1987a; Fowler and Doig 1983; Haynes 1986). These deposits show clearly on airborne gamma-ray spectrometric maps, and the highest grade uranium mineralization is found in veins intruding mafic metavolcanic rocks and mafic volcaniclastic metasedimentary rocks of the Eels Lake Formation. Skarn-type, carbonate-hosted, uranium mineralization is less common in the Harvey–Cardiff Arch than in the adjacent Tory Hill area of the Bancroft Terrane (e.g., Haynes 1986), possibly due to the presence of fewer carbonate units in the arch. It is not clear if molybdenum mineralization shows the same distribution as uranium, although in the Bancroft Terrane, Easton (1990a) noted that the highest grade molybdenite occurrences were found in areas underlain by amphibolites and dolomitic marbles.

Industrial mineral prospects in the Harvey–Cardiff Arch are limited by the lack of suitable host rocks or areas of suitable metamorphic and/or structural histories. The limited abundance of marbles, as well as the diorite-gabbro plutons which produce metamorphic aureoles and/or hydrothermal regimes needed to make suitable industrial mineral prospects, mean that few high-grade industrial



44°48'30"N
78°50'W
Denna Lake Structural Complex
East Moore Lake
Moore Lake
lot 23 - Hamilton alloys
South Beaver Lake
35
thern boundary
y-Lutterworth map area
Horscroft
OGS-1984
503
Shadow Lake
land
Horscroft Mine
Russell-Ponton Mine
0
1 km
78°50'W
44°45'N
78°45'W
C
F
G
H
J
K
Areas most favourable for molybdenum occurrences
Paleozoic rocks
Glamorgan gneiss
Disrupted gneisses
Central Gneiss Belt orthogneiss
C Marble breccia, unsubdivided
F Dolomite marble
G Dolomite marble with amphibolite layers
H Dolomitic, marble, arenite, quartzofeldspathic gneisses
J Granite
K Quartzite, granite, migmatite
CMBBZ Central Metasedimentary Belt Boundary Zone
Mineral occurrence
Geological boundary within Denna Lake Structural Complex
Geological boundary
35 Highway

**Figure 19.76.** Relationship of mineral deposits to "ghost stratigraphy" and tectonic boundaries within tectonically disrupted strata of the Denna Lake Structural Complex in the Minden area (*modified from* Easton 1990a).

mineral prospects are present in the region. Siliceous clastic metasedimentary rocks are predominantly volcanic-derived and, apart from garnet, are not likely to yield aluminous minerals. Metaplutonic rocks in the area are texturally complex, and few stone prospects in the area have yet been identified (cf., LeBaron et al. 1990). The lack of well-defined shear zones in the area means that the flagstone potential in the Harvey–Cardiff Arch is limited.

## Belmont Domain

The Belmont Domain constitutes most of the classic "Hastings Basin" or "Hastings Metamorphic Low" and is probably the most extensively studied part of the Central Metasedimentary Belt because of the concentration of low metamorphic grade supracrustal rocks, easy access and a variety of metallic and industrial mineral deposits. As a consequence, studies in the Belmont Domain have generally

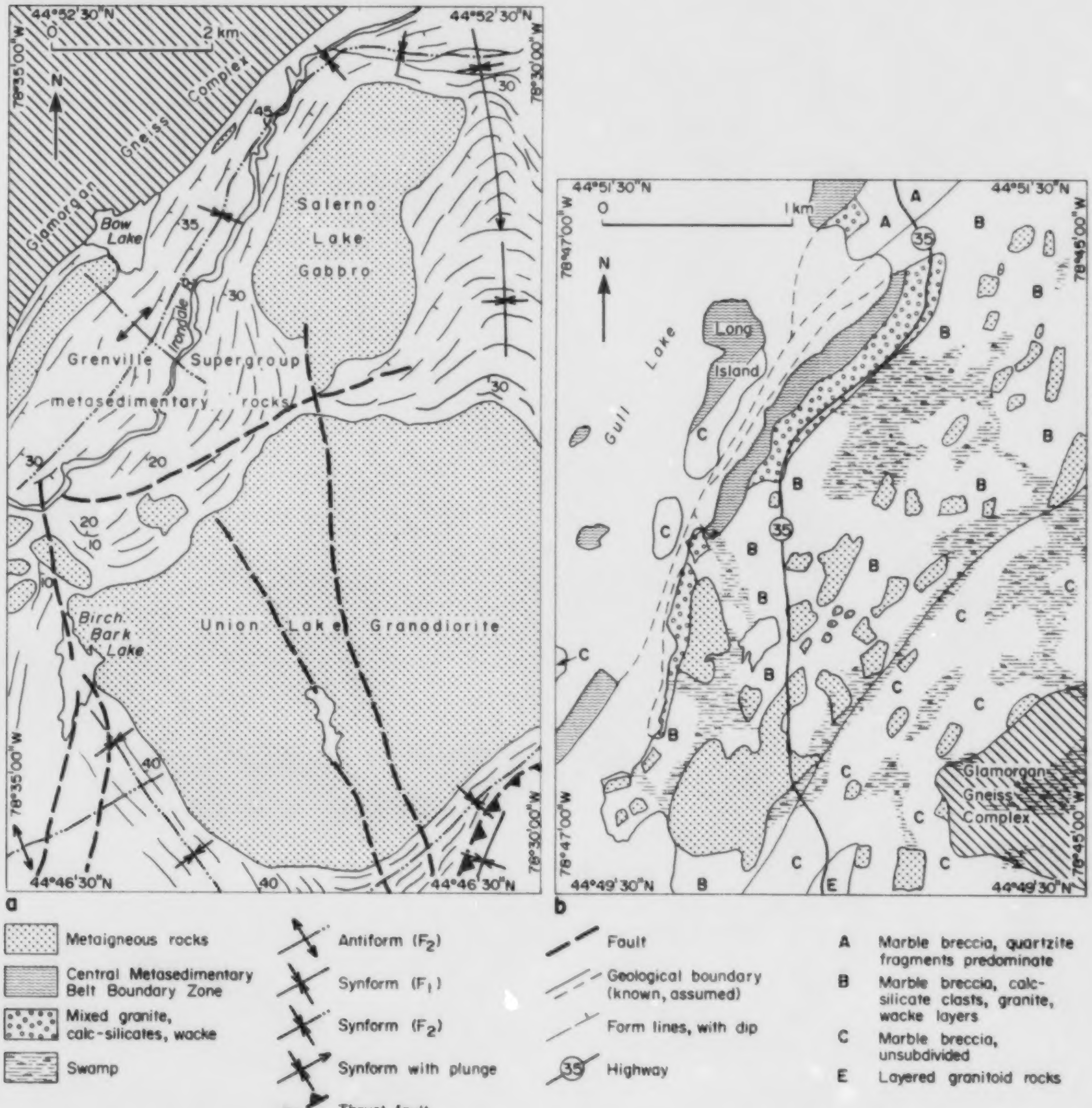

44°52'30"N
0
2 km
N
78°35'00"W
78°30'00"W
Glamorgan Gneiss Complex
Bow Lake
Salerno Lake Gabbro
Irondale R.
Grenville Supergroup metasedimentary rocks
Union Lake Granodiorite
Birch Bark Lake
45
35
30
20
10
40
44°46'30"N
a
44°51'30"N
0
1 km
N
78°47'00"W
78°45'00"W
Long Island
Gull Lake
35
A
B
C
E
Glamorgan Gneiss Complex
44°49'30"N
b
Metaigneous rocks
Central Metasedimentary Belt Boundary Zone
Mixed granite, calc-silicates, wacke
Swamp
Antiform (F2)
Synform (F1)
Synform (F2)
Synform with plunge
Thrust fault
Fault
Geological boundary (known, assumed)
Form lines, with dip
Highway
A Marble breccia, quartzite fragments predominate
B Marble breccia, calc-silicate clasts, granite, wacke layers
C Marble breccia, unsubdivided
E Layered granitoid rocks

**Figure 19.77.** Comparison of outcrop patterns of granitic rocks in (a) the Howland map area southeast of the Glamorgan Gneiss Complex and in (b) the Denna Lake Structural Complex. Units in Denna Lake Structural Complex are the same as in Figure 19.72a. Note the absence of large plutons in the Denna Lake Structural Complex (*modified from* Easton 1990a).

been considered as models applicable to the remainder of the Central Metasedimentary Belt.

The domain has been extensively mapped, with most of the domain covered by 1:31 360 scale mapping or better (Hewitt and James 1956; Hewitt and Satterly 1957; Hewitt 1961, 1964a, 1968; Lumbers 1964, 1967a, 1968, 1969; Laasko 1968; Bartlett and Moore 1985; Di Prisco 1989). Bartlett and deKemp (1987) show the distribution of carbonate units, stromatolites and carbonate-hosted mineral deposits in the domain. Numerous geochronologic studies from the Belmont Domain are summarized in Easton (1986a). Table 19.1 lists metamorphic studies conducted within the domain.

## STRATIGRAPHIC RELATIONSHIPS

Lumbers (1967a, 1967b) established the stratigraphic framework of the Belmont Domain (and the Central Metasedimentary Belt) shown in Figure 19.53b, building on the work of Hewitt and James (1956). Some fine tuning of the sequence has occurred, with Bartlett (1983) and Bartlett and Moore (1985) introducing additional formations in the Belmont Lake area (*see* Table 19.12) and Easton (1986d) establishing that the Apsley Formation (Mayo Group) was probably stratigraphically equivalent to the Vansickle Formation, and that, therefore, the Hermon and Mayo groups were laterally interfingering (*see* Figure 19.56). Table 19.12 is a modified version of the stratigraphic sequence for the domain showing these changes.

Figures 19.56, 19.63, 19.64 and 19.70 illustrate the facies relationships within the domain. The environment can be viewed as a series of volcanic islands that developed either as an archipelago or coalesced to form a seaward-facing landmass. These locally emergent volcanic islands produced a variety of volcanic products. Volcanic detritus was quickly eroded off these islands forming large volcaniclastic aprons, generally deposited in turbidite fan systems. Carbonates were deposited in 2 regimes: 1) as shallow-water platforms adjacent to the volcanic centres (e.g., Marmora Formation); or 2) as carbonate turbidites deposited distal to the volcanic centres, but which interfingered with the volcaniclastic detritus of the volcanoes (e.g., Dungannon Formation). The source region

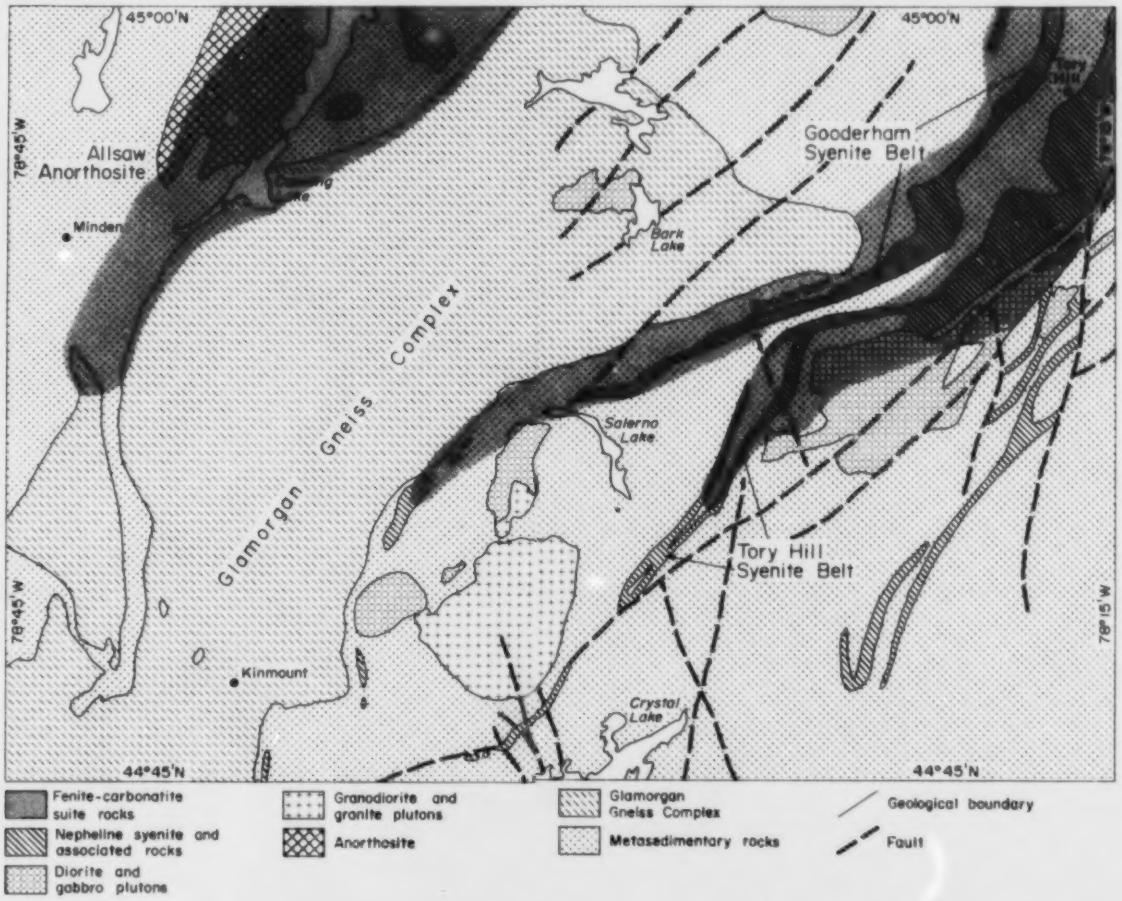

45°00'N
45°00'N
78°45'W
78°45'W
78°15'W
44°45'N
44°45'N
Allsaw Anorthosite
Minden
Bark Lake
Gooderham Syenite Belt
Tory Hill
Glamorgan Gneiss Complex
Salerno Lake
Tory Hill Syenite Belt
Kinmount
Crystal Lake
Fenite-carbonatite suite rocks
Nepheline syenite and associated rocks
Diorite and gabbro plutons
Granodiorite and granite plutons
Anorthosite
Glamorgan Gneiss Complex
Metasedimentary rocks
Geological boundary
Fault

**Figure 19.78.** Distribution of nepheline syenite suite rocks in the Bancroft Terrane. Distribution of fenite-carbonatite suite rocks is also shown for comparative purposes.

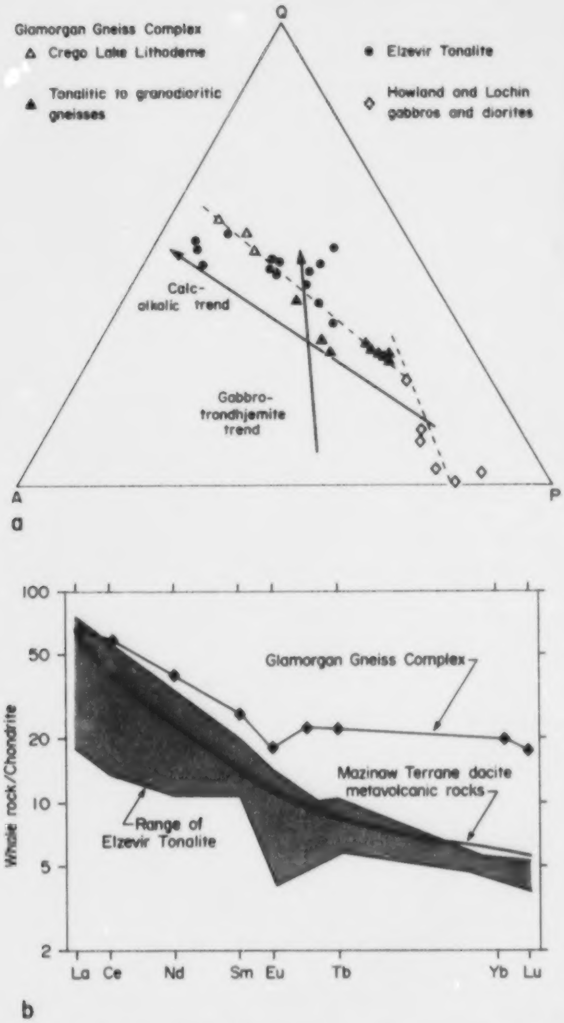


**Figure 19.79.** Comparison of geochemistry of the Glamorgan Gneiss Complex with Elzevir Suite plutons from the Elzevir Terrane. **a)** QAP plot showing similarity between the Glamorgan Gneiss Complex and the Elzevir Tonalite (*modified from* Easton 1987e). **b)** Chondrite-normalized REE patterns showing distinct pattern of the Glamorgan Gneiss Complex representative sample (*see* Easton 1987e for full analyses) compared to the Elzevir Tonalite (data *from* Pride and Moore 1983) and a Mazinaw Terrane dacitic metavolcanic rock (data *from* Condie and Moore 1977).

**Figure 19.80.** Clinopyroxene skarn assemblage in calcitic marbles, Denna Lake Structural Complex, near Minden. Hammer handle is 30 cm long.

of these carbonate turbidites is not preserved within the Belmont Domain. It may have been a shelf region similar to the Bancroft Terrane. Figures 19.60b to 19.60d, 19.61a to 19.61d, 19.66b to 19.66d and 19.67a to 19.67e illustrate a variety of metavolcanic and metasedimentary rocks from the Belmont Domain.

## GEOCHEMISTRY OF THE METAVOLCANIC ROCKS

Volcanic sequences studied include the Turriff metavolcanics (Holm et al. 1985, 1986; Smith and Holm 1987), the Tudor Formation in Tudor Township in the Jordan Lake area (Holm et al. 1986; Smith and Holm 1987) and dike rocks from the Gilmour–Turriff area in the eastern part of the domain (Holm et al. 1986; Smith and Holm 1987) and dike rocks from the western part of the domain (Smith and Holm 1990a). Bartlett (1983) studied the Belmont Lake volcanics. Bright (1988) reports analyses from volcanic rocks from the Stoney Lake area in the western part of the domain.

Smith and Holm (1990a, 1990b) found 4 major types of tholeiites within the dike rocks and among the volcanic successions, namely: 1) a low zirconium suite of basalt and minor dacite; 2) an intermediate zircon suite of basalts, andesites, dacites and rhyolites (e.g., Tudor Township volcanic rocks); 3) a high zirconium suite of high-iron basalts; and 4) highly fractionated basalts with high Mg-number, high nickel and chromium, with strongly fractionated trace-element patterns represented by only a few basaltic dikes. Also present in the dikes were 5) a calc-alkalic suite of basalts and andesites; 6) a transitional suite of andesites and trachyandesites; and 7) a single mafic dike of ultrapotassic composition. The latter may be genetically related to the syenite plutonic suite. Holm et al. (1985, 1986) report that the Turriff metavolcanics are a bimodal suite composed of tholeiitic basalts and dacite-rhyolite pyroclastic rocks, with andesites in the sequence showing calc-alkalic affinities.

Bartlett (1983) reported a similar sequence in the Belmont Lake area, with a bimodal sequence of tholeiitic basalts and dacite and rhyolite, with calc-alkalic andesites occurring above the basalts but below the dacites and rhyolites. In the Tudor area, Holm et al. (1986) found only a bimodal sequence of tholeiitic basalt and dacite and rhyolite. Holm et al. (1986) and Smith and Holm (1987, 1990a) suggest a tectonic environment of attenuated continental crust in an ensialic back-arc basin, based on geochemistry and the plutonic and sedimentary rocks associated with the volcanic sequence. Bartlett (1983) also favoured an ensialic back-arc basin model.

Bright (1988) analyzed volcanic rocks in the Stoney Lake area and documents a sequence of dominantly tholeiitic basalts and a calc-alkalic sequence of andesite-dacite-rhyolite. This is consistent with geochemical data *of* Easton (1989e) (*see* Figure 19.65a to e) from the Apsley Formation volcaniclastic rocks indicating a calc-alkalic source.

## PLUTONISM

Plutons of the nepheline syenite, Elzevir, Methuen, Lavant and Skootamatta suites occur in the Belmont Domain, with Methuen and Lavant intrusions predominating. Most plutons have ovoid to slightly elongate forms and contact aureoles (e.g., Deloro Granite, Cordova Gabbro).

The Blue Mountain Nepheline Syenite (Derry 1951; Phipps 1955; Derry and Phipps 1957; Hewitt 1961; Payne 1966, 1968) is the only representative of the nepheline syenite suite in the Elzevir Terrane. The main body is 2 by 4 km, and a long sill-like arm about 400 m wide extends 6 km southwest. Both intrude volcaniclastic metasedimentary rocks and marbles. Krogh and Hurley (1968) report a whole-rock Rb-Sr age of 1258±41 Ma from the body. The Blue Mountain Nepheline Syenite is an important industrial mineral operation, with commercial exploitation beginning in 1936 and continuing to the present (*see* Easton et al. 1986; MacGregor 1983). The nepheline is used primarily within the ceramic industry.

Numerous diorite-gabbro intrusions are present in the domain, including the Thanet, Umfraville, Tudor, Methuen, Cordova and Horseshoe Lake gabbros and the Tallan Lake Sill. These intrusions consist of a variety of phases, including diorite, gabbro, pyroxenite and anorthositic gabbro and may be cut by late tonalite and syenite dikes. Lumbers (1969) describes the Tudor, Thanet, Jocko Lake and Umfraville gabbros in detail. Chemical data on some of these bodies is presented in Lumbers (1969) and Lumbers et al. (1990). U-Pb zircon ages indicate that these bodies are *circa* 1245 to 1240 million years in age (Davis and Bartlett 1988; Lumbers et al. 1990).

The predominant intrusive rock in the Belmont Domain is the alaskitic granites of the Methuen Suite. These include the Methuen, Coe Hill and Deloro granites. Saha (1959) studied the petrography of the Deloro Granite, as did Abdel-Raham and Martin (1987). Lumbers et al. (1990) summarize the chemistry of these bodies and classify them as anorogenic granites. U-Pb zircon dates (van Breemen and Davidson 1988b; Lumbers et al. 1990) indicate the granites were emplaced at *circa* 1245 to 1240 Ma, coeval with the Lavant Suite gabbros. The genetic relationship between these 2 magmatic events is not known. The Deloro Granite is characterized by higher fluorine contents than other plutons of the suite (Wu 1984; Wu and Kerrich 1986). The Deloro Granite has been affected by a low temperature isotopic resetting event of Ordovician age (Lopez-Martinez and York 1983), which reset both potassium-argon and rubidium-strontium isotopic systems. This isotopic resetting event may be associated with barite-fluorite-calcite vein emplacement in the Madoc area.

Skootamatta Suite plutons in the domain include the Gawley Creek Syenite and the Chandos (Loon Lake) Syenite (Dostal 1975; Heaman 1985). These plutons show the same ultrapotassic chemistry as other members of the suite, and the characteristic aeromagnetic high associated with members of this suite.

## STRUCTURAL HISTORY

Structural studies have focussed on the northwestern part of the domain near the Elzevir–Bancroft terrane boundary (Best 1966), or the Harvey–Cardiff Arch boundary (Jennings 1969). In general, these studies revealed that the Belmont Domain is characterized by open, nearly recumbent folds that have gentle plunges. Considerably more work needs to be done to characterize the structural history of the Belmont Domain and to compare this history with adjacent regions.

## METAMORPHIC HISTORY

The eastern part of the domain lies within the "Hastings Metamorphic Low" and is characterized by middle to upper greenschist-facies metamorphic assemblages (Lumbers 1964, 1967a, 1969), although amphibolite-facies assemblages have been noted adjacent to the Deloro Granite (Sampson 1972; Sampson and Fawcett 1977). The western part of the domain is preserved at lower to middle amphibolite facies. Table 19.1 summarizes the metamorphic studies that have been conducted within the domain.

The domain was subjected to 1 regional metamorphic event at *circa* 1230 to 1180 Ma (Lopez-Martinez and York 1983; Lumbers et al. 1990), corresponding to the Elzevirian Orogeny. As illustrated in Figure 19.16, the argon-argon cooling curve of the domain after this event was perturbed a second time, at *circa* 1140 Ma. This event may correspond to the onset of the Ottawan Orogeny.

## MINERAL DEPOSITS

The most significant mineral deposit type is the sulphide deposits and iron formations associated with volcanic centres in the domain, magmatic iron-titanium and copper-nickel deposits and contact metasomatic deposits including iron, wollastonite and talc.

### Sulphide Deposits and Iron Formation

These deposits are associated with volcanic centres in the domain, near volcanic-sediment contacts. The most extensive of these deposits are associated with the Belmont Lake volcanics (Carter 1984; Bartlett and Moore 1985) and consist of rusty schists containing gold, silver, copper, nickel, cobalt and zinc (e.g., Deer Lake schist, Carter 1984, Easton et al. 1986), as well as magnetite-chert iron formation. The iron formation units are thin, and of limited lateral extent (less than 400 m; Carter 1984).

The Canadian Sulphur Ore and Blakely pyrite deposits produced pyrite for use in sulphuric acid production. The stratigraphic setting of these deposits is unclear (Carter 1984; Verschuren 1984). They are not directly associated with a volcanic centre and are associated with clastic metasedimentary rocks and marbles that have been altered by the Deloro Granite (Easton 1989a; H. Meyn, Ministry of Northern Development and Mines, Bancroft, personal communication 1990, 1991). Altered mafic volcanic rocks may be present within the sequence. No other sizeable sulphide deposits have been reported from the domain.

#### Magmatic Sulphide and Iron-Titanium Deposits

Copper-nickel sulphide deposits are associated with Lavant Suite plutons and include the Macassa deposit (Lumbers 1969; Carter 1984). Iron-titanium deposits are also associated with Lavant Suite plutons, the largest of which is the Methuen ilmenite deposit (Scoates et al. 1988).

#### Contact Metasomatic Deposits

The Marmoraton iron mine, which produced roughly 25 million tons of iron ore between 1955 and 1978 (*see* summary *in* Carter 1984; Easton et al. 1986), is the largest and most significant of the metallic iron deposits. Numerous others are located adjacent to the Deloro Granite and other Methuen and Lavant suite plutons in the domain. Near the Paleozoic unconformity, the upper 50 to 100 m of these magnetite deposits are commonly altered to hematite (Di Prisco and Springer 1991).

Industrial minerals also occur in suitable host rocks in the contact aureoles of plutons within the domain and include wollastonite (MacKinnon 1990) and talc (Canada Talc Mine, *see* summary *in* Easton et al. 1986).

#### Gold Deposits

Epigenetic quartz and quartz-ankerite vein deposits occur throughout the greenschist facies part of the Belmont Domain (Carter 1984; LeBaron 1991). The Malone gold deposit occurs in a sheared satellite intrusion of the Deloro Granite and is associated with the intersection of 2 regional shear zones (LeBaron 1991). This low-grade, large tonnage deposit is distinct from the typical gold vein deposits described elsewhere in the Central Metasedimentary Belt (*see* Easton and Fyon, this volume).

#### Industrial Minerals

A variety of industrial minerals are present within the Belmont Domain. Contact metasomatic talc and wollastonite deposits have already been noted. Dolomite marbles in the Madoc area are quarried by Stokoslar for use as roofing granules, and porcelanities surrounding the Deloro Granite have also been examined for similar use (Easton 1989a). Building stone sites in the area are summarized by LeBaron et al. 1989, 1990. Dolomite resources in the area are summarized by LeBaron and MacKinnon (1990).

## Grimsthorpe Domain

The Grimsthorpe Domain is one of the least complex Elzevirian domains and has recently been mapped by Easton and Ford (1990a, 1990b) and Di Prisco (1989). It is composed of about 65% plutonic rocks, with supracrustal rocks comprising the remainder. The stratigraphic succession consists of older tholeiitic flows and mafic intrusions (Canniff Complex) overlain by a volcaniclastic sequence (Grimsthorpe Group), which has been intruded by a younger ultramafic-mafic plutonic suite, in turn intruded by tonalite plutons of the Elzevir Suite (*see* Table 19.13). Syenites of the Skootamatta Suite are also common within the domain. The most deformed rocks in the domain are the older Canniff Complex. The volcaniclastic sequence (Grimsthorpe Group) and the gabbro and tonalite bodies preserve primary textures, including igneous layering. Plutons show semicircular to oblate forms approximating original intrusive forms. Metamorphism varies from middle greenschist to upper amphibolite. Isotopic resetting of Rb-Sr systems in the area is incomplete (e.g., Rb-Sr age [1240±50 Ma] on the Elzevir Tonalite is only slightly younger than the zircon age of 1270 Ma).

### GEOCHEMISTRY OF THE METAVOLCANIC ROCKS

This succession has been sampled for geochemistry by Di Prisco (1989), LeBaron and van Haaften (1989) and Easton and Ford (1990a, in prep.). Preliminary results of these studies indicate that only tholeiitic basalts are present. In addition, volcaniclastic metasedimentary rocks of the Grimsthorpe Group are basaltic to basaltic andesite in composition, reflecting derivation from a dominantly tholeiitic basaltic or gabbroic source region. The volcaniclastic metasedimentary rocks characteristically have $Na_2O$ contents of 3.5 to 4.5%, whereas metawackes and semipelites locally interbedded with the volcaniclastic metasedimentary rocks have $Na_2O$ contents of less than 2.5%, a reflection of the mafic source region for the volcaniclastic metasedimentary rocks.

### PLUTONISM

The domain is underlain by 4 plutonic suites, with tonalitic rocks of the Elzevir Suite predominant. The oldest rocks comprise mafic to ultramafic, protomylonitic mafic intrusive rocks (Figure 19.81) preserved in a thin band found

**Figure 19.81.** Protomylonitic gabbros of the Canniff Complex, Grimsthorpe Domain. Hammer handle is 30 cm long.

along the western margin of the Elzevir Tonalite along both margins of the Canniff Tonalite (Easton and Ford 1990a, 1990b, in prep.) and along the southwestern and west-central margin of the Weslemkoon Tonalite. These metagabbros and related rocks (Canniff Complex, *see* Table 19.13) shed clasts into mafic volcaniclastic metasedimentary rocks within the domain (*see* Figure 19.60e). The ultramafic portions of these bodies are locally altered to tremolite-talc-carbonate, as are some of the conglomerates containing abundant clasts of these ultramafic rocks. Chappell (1978) noted similar relationships within the Kaladar Complex on the eastern margin of the Elzevir Tonalite and distinguished these older ultramafic metaplutonic rocks from younger hornblendites of Killer Creek Suite age. Chappell (1978) suggested that these metaplutonic rocks were among the oldest rocks in the Central Metasedimentary Belt and served as a basement to the "Tudor metavolcanics." Thus, the Canniff Complex and the older ultramafic sequence (serpentinites, talc schists) of the Kaladar Complex may represent an ophiolitic basement to the Grimsthorpe (and Hermon?) Group (Easton and Ford, in prep.; Brown et al. 1975). Major and trace element geochemistry of the Canniff Complex rocks (Easton and Ford, in prep.) indicates an ocean floor basalt affinity.

LeBaron and van Haaften (1989), LeBaron (1988), LeBaron et al. (1987) and Di Prisco (1989) show the distribution of the tremolite-talc occurrences associated with the Canniff Complex. The tremolite-talc occurrences are compositionally equivalent to basaltic and peridotitic komatiites, although texturally these rocks are clearly intrusive in origin (Easton and Ford 1990a, 1991, Di Prisco 1989).

Synvolcanic metagabbros are present throughout the domain, as lenticular intrusions cutting the supracrustal rocks. A variety of metagabbroic dikes are also present throughout the supracrustal sequence.

The Killer Creek Gabbro is a large pluton of medium- to coarse-grained metagabbro with lesser melanocratic metagabbro, anorthositic gabbro and hornblendite, locally, with magnetite-chromite-bearing horizons (Easton and Ford 1990a, 1990b). Igneous layering is locally well-preserved. The body has been intruded by the *circa* 1267 million-year-old Elzevir Tonalite (Heaman, *in* Davidson et al. 1990), and spectacular intrusion breccia zones are present where the 2 are in contact. A large aeromagnetic anomaly located northeast of Joeperry Lake in Bon Echo Provincial Park corresponds to a zone of altered melanocratic metagabbro and hornblendites. Other parts of the Killer Creek Suite do not have a notable aeromagnetic signature. Preliminary geochemical data on the body shows a calc-alkalic to alkalic trend, with elevated $TiO_2$ (greater than 2.5%) and $P_2O_5$ (greater than 0.6%).

Tonalitic to granodioritic rocks of the Elzevir Suite underlie most of the domain and include the Elzevir, Canniff and Weslemkoon tonalites and the Lingham Lake Gabbro-Diorite Pluton. Supracrustal rocks are preserved in narrow bands between the plutons, and it is likely that the plutons are continuous at depth. Gravity modelling suggests the plutons extend to a depth of 3 to 5 km (Real and Thomas 1987).

The Elzevir Tonalite may be a composite pluton, based on changes in grain size and textures between the northern and central parts of the body and the distribution of xenoliths (Easton and Ford 1990a, 1990b). The Lingham Lake Gabbro-Diorite Pluton (Lingham Lake Complex *of* Lumbers 1969) consists mainly of medium-grained quartz diorite with numerous schlieren-rich and mafic xenolith-rich zones that has intruded an older gabbroic body similar to the synvolcanic gabbroic rocks present in the domain. The older gabbroic rocks, for mapping purposes, are included as part of the suite. It is probable that the Lingham Lake Gabbro-Diorite Pluton represents a high-level cross section through one of the tonalitic plutons, with the slightly more mafic composition the result of incorporation of mafic host rocks.

The geochemistry of the Elzevir Tonalite has been studied by Pride and Moore (1983) who classified it as a high-alumina tonalite that probably formed in an island-arc environment. They also suggested a genetic link between Mazinaw Group dacites and Elzevir Suite tonalites, with both being derived from the melting of lower crustal material.

Two plutons of the Skootamatta Suite are present in the domain, the Mount Moriah Syenite and the Skootamatta Syenite. The Mount Moriah Syenite consists of medium- grained, grey, hornblende biotite syenite and coarse-grained, pink, biotite syenite. Lamprophyres are locally present, particularly along the margin of the pluton. The Skootamatta Syenite consists of coarse-grained, grey, biotite monzonite and pink, coarse-grained biotite syenite. Ultrapotassic lamprophyric dikes are locally present along the margin of the body but are less abundant then adjacent to the Mount Moriah Syenite (Easton and Ford, in prep.). Both bodies are characterized by strong aeromagnetic signatures which correspond closely to the outcrop exposure of the plutons.

## STRUCTURAL HISTORY

The structural history of the domain is relatively simple. Few if any regional fold structures have been identified, and most rocks in the domain show little strain. Conglomerates preserve well-rounded cobbles and small-scale sedimentary textures, including dewatering textures. Way-up features are locally widely distributed in metavolcanic and metasedimentary rocks in the domain. In contrast, the older mafic and ultramafic intrusive rocks, as well as the older mafic metavolcanic rocks (i.e., Canniff Complex), show higher degrees of strain and are locally protomylonitic, particularly the intrusive rocks. Bedding parallel foliation is developed in most supracrustal rock units, and in the west-central part of the domain, a second east-trending cleavage and late shears are present. Mafic dikes of several ages cut the supracrustal units and the Weslemkoon Tonalite. Most dikes are near vertical and strike 060°.

A deformation zone, up to 100 m wide, the Partridge Creek shear zone, transects the central part of the domain and separates the Elzevir Tonalite and the Killer Creek

Gabbro in the east from the supracrustal and metaplutonic rocks of the western part of the domain (Easton and Ford 1990a, 1990b). The eastern part of the domain may expose a slightly deeper level of exposure than the western part of the domain.

The Mooroton shear zone marks the eastern margin of the Grimsthorpe Domain with the Mazinaw Terrane. South of Flinton, the zone is narrow where it separates a thin sliver of Flinton Group metasedimentary rocks and minor mafic metavolcanic rocks and ultramafic pods from the Northbrook Tonalite of the Mazinaw Terrane. Along the Skootamatta River between Northbrook and Skootamatta Lake, a westerly zone of straight gneisses, about 150 m in width, containing deformed pods of Killer Creek gabbroic rocks is present. East of the terrane boundary is a zone, up to 1 km wide, of chloritized, sheared mafic metavolcanic rocks, locally containing gold-arsenopyrite mineralization disseminated in the metavolcanic rocks and in quartz vein systems (e.g., Addington Mine, O'Donnell occurrences; Wolff 1982a; Dillon 1985; Moore and Morton 1986; Laing et al. 1987; Harnois and Moore 1989). Farther north, the zone is ill-defined where it is truncated by north-northwest-trending faults (Easton and Ford 1991).

In the west, Di Prisco (1989) identified a lithologic discontinuity between carbonate metasedimentary rocks of the Belmont Domain and deformed metaigneous rocks of the Grimsthorpe Domain. Gold mineralization hosted in quartz and quartz-carbonate veins is present along this discontinuity (cf., LeBaron 1991). North of Bannockburn, the domain boundary has not been mapped in detail but probably lies at the marble-metavolcanic contact.

Connelly (1985, 1986) conducted a detailed structural study of the margin of the Elzevir Tonalite. He noted a discordance in lineations between the Elzevir and Northbrook tonalites and suggested it may represent a major tectonic break. This break corresponds to the southern extension of the Mooroton shear zone.

## METAMORPHIC HISTORY

The protomylonitic mafic and ultramafic intrusive rocks and the Canniff Complex may have been subjected to a pre-1250 million-year-old period of low-grade metamorphism, probably occurring soon after emplacement. The bulk of the rocks in the Grimsthorpe Domain, including all the metaigneous rocks, except the Skootamatta and Mount Moriah syenites, have been subjected to a regional metamorphic event of unknown age. This event could have occurred at 2 main times, *circa* 1180 to 1165 Ma (same age as Frontenac Terrane metamorphism) as suggested by Lumbers et al. (1990), or *circa* 1100 to 1070 Ma, coincident with deformation in the Central Gneiss Belt and titanite ages in the Belmont Domain (Davis and Bartlett 1988; Lumbers et al. 1990). K-Ar hornblende and muscovite ages indicate cooling below blocking temperatures of 2 respective minerals at *circa* 1020 Ma (Hanes et al. 1988; Cosca 1989; Figure 19.17). The southern half of the Grimsthorpe Domain was subjected to greenschist-facies conditions, rising to lower amphibolite facies in the north.

Contact metamorphism may also have occurred coincident with emplacement of the tonalite-granodiorite plutons (Lumbers 1968, 1969), although the contact aureoles appear to be narrow. The extent of the contact metamorphic event is difficult to define, in part, because the Canniff Complex metaigneous rocks that were subjected to an earlier metamorphic event are located in close proximity to the tonalite plutons.

Wu et al. (1991) compared the chemistry of the low- and high-grade parts of the Weslemkoon Tonalite. Rubidium, strontium and barium showed greater dispersal in the high-grade rocks, suggesting local remobilization. Total REE contents were lower in the high-grade rocks (though the patterns remained unchanged), an effect Wu et al. (1991) attributed to titanite breakdown.

## BASEMENT

The Canniff Complex exposed near Lingham Lake, along the western flank of the Canniff and Weslemkloon tonalites and in the Kaladar Complex, served as basement to the volcaniclastic sequence. It is possible that the Canniff Complex represents a sliver of oceanic crust.

## MINERAL DEPOSITS

Few metallic mineral deposits have been reported from the domain, the core of which was poorly accessible until recently. Gold deposits, hosted in early shear zones, locally reconcentrated in crosscutting vein systems (e.g., Gilmour Mine, Lumbers 1968, Carter 1984, Christie 1992) are present mainly along the margins of the Grimsthorpe Domain. These deposits are also generally coincident with the greenschist-amphibolite facies transition.

A few small gold occurrences are located within the central part of the domain, hosted in mafic volcaniclastic metasedimentary rocks but are generally small and of limited extent (e.g., United Reef Petroleums Property, Easton and Ford 1990a, in prep.).

Rusty schist horizons, representing sulphidic metamorphosed black shales and metasiltstones, in part, sulphide-facies iron formation, occur throughout the Grimsthorpe Group. To date, no significant mineralization has been identified from these units (Easton and Ford 1990a, in prep.). A zone of albitization and silicification, associated with an aeromagnetic anomaly, is located within the central part of the Grimsthorpe Domain in Angelsea Township (Easton and Ford 1990a, in prep.), however, no significant mineralization has been reported from this zone.

Easton and Ford (1990a, in prep.) report local concentrations of chromite and magnetite and possibly ilmenite, as continuous horizons roughly coincident with igneous layering in hornblendites within the Killer Creek Gabbro. Disseminations of chalcopyrite are common throughout the gabbro. This type of mineralization has only recently been identified and has yet to be thoroughly prospected.

As noted above, older, protomylonitic ultramafic intrusive rocks contain numerous tremolite-talc bodies. These occurrences and their potential are described in detail by LeBaron and van Haaften (1989).

Several building stone prospects have been identified from the Grimsthorpe Domain, all located within the intrusive rocks, including the Elzevir and Weslemkoon tonalites, the Skootamatta and Mount Moriah syenites and the Killer Creek Gabbro. These are described by LeBaron et al. (1990).

## MAZINAW TERRANE

The Mazinaw Terrane (*see* Figures 19.51 and 19.82) is characterized by: 1) a linear structural pattern of northeast-trending, alternating, metaplutonic and supracrustal belts (*see* Figure 19.82); 2) higher metamorphic grades than either the adjacent Elzevir or Sharbot Lake terranes; 3) a polycyclic metamorphic and structural history; 4) a slower cooling from peak metamorphic conditions, which has resulted in more complete isotopic resetting of rubidium-strontium and potassium-argon isotopic systems (data *in* Cosca 1989; *see* Figures 19.16 and 19.17); and 5) the presence of the Flinton Group, a metamorphosed supracrustal sequence that unconformably overlies the Mazinaw Group (Grenville Supergroup). Metasedimentary rocks underlie about 25% of the terrane, metavolcanic rocks about 20% and metaplutonic rocks 55%. Although many mapping studies (Smith 1958; Smith et al. 1969; Wolff 1982a, 1982b; Bright 1986a; Moore and Morton 1986; Pauk 1987, 1989a, 1989b; Easton 1988a, 1990e) have been conducted in the Mazinaw Terrane, only recently has the complex history and distinct nature of this terrane begun to be fully understood.

Previous workers considered the Mazinaw Terrane to be part of the Elzevir Terrane. The time-space chart for the Mazinaw Terrane (Table 19.17) shows that the terrane has had a more complex history than the adjacent Elzevir and Sharbot Lake terranes, as well as a different structural style and magmatic history. In some respects, the Mazinaw Terrane provides a link between the Elzevir and Frontenac terranes. Tonalites, similar to those of the Grimsthorpe Domain, may have served as basement to the supracrustal succession in the Mazinaw Terrane. The Addington Granite appears to represent Methuen alaskitic magmatism seen in the Elzevir Terrane, although the granitic bodies within the Mazinaw Terrane are more heterogeneous in composition and include a high sedimentary xenolith component not typical of anorogenic granites. The structural style of the Mazinaw Terrane, however, is similar to that observed in the Frontenac Terrane and Adirondack Lowlands, and its polycyclic metamorphism seems more closely linked to Frontenac Terrane, not Elzevir Terrane, metamorphism. Finally, the age of detrital zircons in the Flinton Group ties the Mazinaw Terrane in with Frontenac Terrane magmatism.

### STRATIGRAPHIC RELATIONSHIPS

#### Mazinaw Group

Most supracrustal rocks in the Mazinaw Terrane are assigned to the Mazinaw Group (new term). Two main belts of supracrustal rocks are present (*see* Figure 19.82), one located in the Kaladar area, including the Clare River synform, the other stretching from Bishop Corners northeast to Clyde Forks through Fernleigh, Ardoch and Ompah. Metavolcanic rocks of mafic, intermediate and felsic composition are present in both belts and are described in greater detail below, under the section on volcanic geochemistry. Both belts include mafic to intermediate hornblende-bearing gneisses ("para-amphibolites"), which may represent interbedded metavolcanic tuffs and metasedimentary rocks of volcanic provenance (e.g., Wolff 1982a; Pauk 1987; Easton and Ford 1991). Siliceous clastic metasedimentary rocks consisting of metawackes, metalitharenites and metasiltstones are common in both belts, as are dolomitic and calcitic marbles. Stromatolites have been reported in dolomitic marbles at Marble Lake (Moore and Morton 1986). Primary textures and structures are not well-preserved making stratigraphic correlation and identification of depositional environments difficult. In both supracrustal belts, it appears that the volcanic edifices had associated aprons of volcaniclastic debris (para-amphibolites and metawackes) interfingered with clastic sediments. Carbonate sedimentation occurred coincident with clastic sedimentation, but the bulk of the carbonate rocks may have been deposited after volcanism and associated sedimentation. In the Clare River synform, clastic and carbonate metasedimentary rocks become more abundant to the east. No systematic variation is readily apparent in the northern belt, although major accumulations of metavolcanic rocks seem to be located mainly in the Harlowe and Flower Station areas, with clastic and carbonate metasedimentary rocks predominant in the intervening Plevna–Ompah area.

#### Flinton Group

The Flinton Group is a distinctive package of metaconglomerate, metamorphosed quartz arenite and metapelite. Ambrose and Burns (1956) introduced the term Flinton Group for some (but not all) conglomeratic rocks in the Kaladar area, but it was not until the 1970s that Moore and Thompson (1972, 1980) formalized the stratigraphy of this sequence of metasedimentary rocks by defining the Flinton Group and several distinctive formations (*see* Table 19.14). Moore and Thompson (1972, 1980) also noted the presence of carbonate within the Flinton Group, namely, the Myer Cave, Lessard and Fernleigh formations. Figures 19.83a to 19.83f, 19.84, 19.85a and 19.85b illustrate the variety of rock types present within the Flinton Group. Stromatolites have not been reported from the Myer Cave Formation, although Bright (1986a) reports stromatolites from the Flinton Group in the Clare River area (*see* Figure 85b).

Table 19.14 summarizes the stratigraphy of the Flinton Group, as modified by Ford (in prep.) and Easton and Ford (1991). Chappell (1978) proposed 2 additional formations in the Clare River area, the Beatty and Bogart formations which are roughly correlative with the Bishop Corners and Lessard formations (*see* Table 19.14). Bright (1986a) mapped the Flinton Group in the Clare River area in detail but did not use Chappell's stratigraphy. Hounslow and Moore (1967) present several major element analyses of Flinton Group schists from the Marble Lake–Fernleigh Belt, however, they are not tied to stratigraphy and appear to

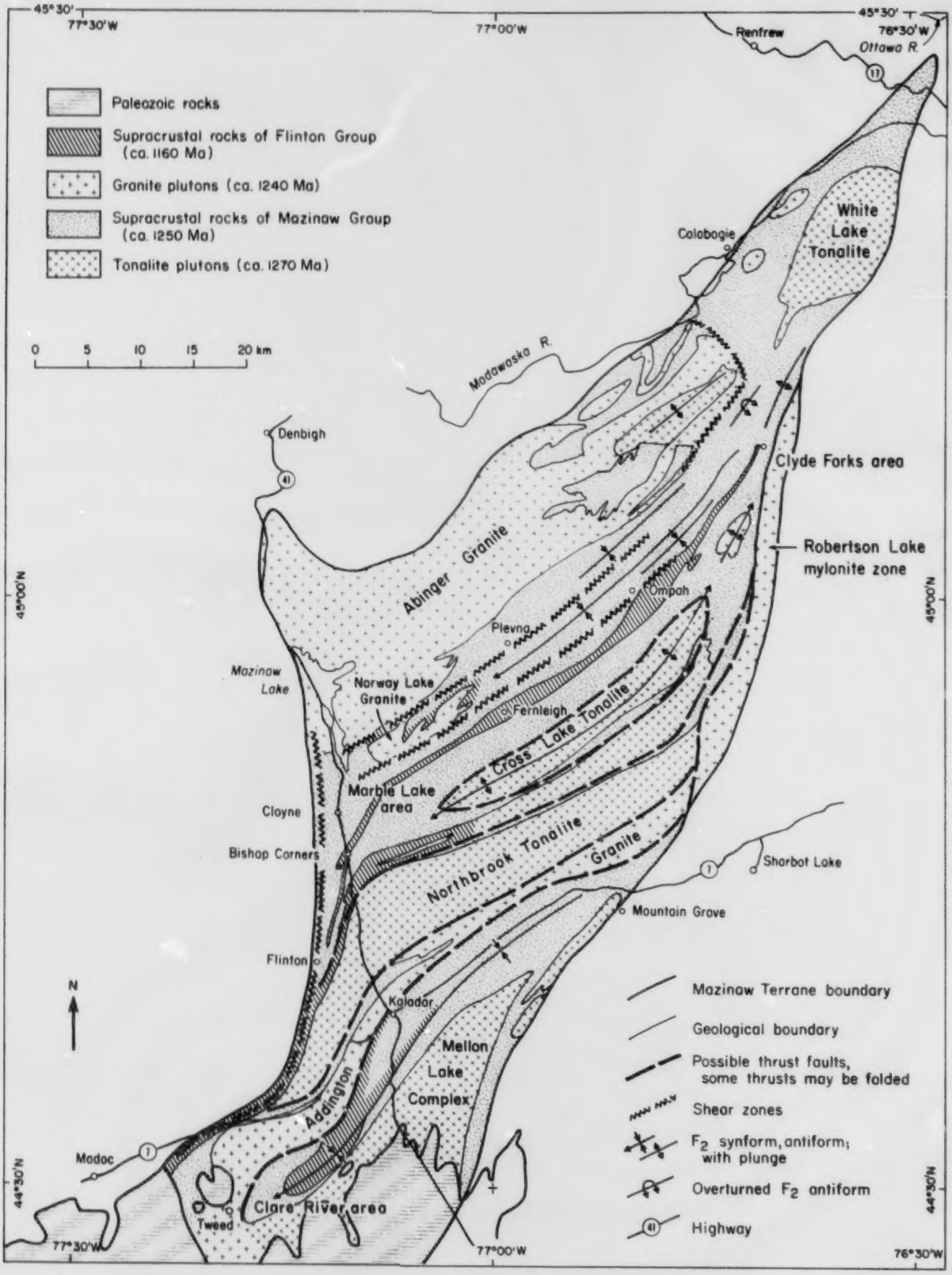

45°30'
77°30'W
77°00'W
Renfrew
45°30'
76°30'W
Ottawa R.
17
Paleozoic rocks
Supracrustal rocks of Flinton Group (ca. 1160 Ma)
Granite plutons (ca. 1240 Ma)
Supracrustal rocks of Mazinaw Group (ca. 1250 Ma)
Tonalite plutons (ca. 1270 Ma)
0 5 10 15 20 km
Calabogie
White Lake Tonalite
Madawaska R.
Denbigh
41
Clyde Forks area
Robertson Lake mylonite zone
Abinger Granite
Ompah
Plevna
45°00'N
Mazinaw Lake
Norway Lake Granite
Fernleigh
Cross Lake Tonalite
Marble Lake area
Cloyne
Bishop Corners
Northbrook Tonalite
Granite
7
Sharbot Lake
Mountain Grove
Flinton
Kaladar
N
Mellon Lake Complex
Addington
Madoc
7
44°30'N
Tweed
Clare River area
77°30'W
77°00'W
76°30'W
Mazinaw Terrane boundary
Geological boundary
Possible thrust faults, some thrusts may be folded
Shear zones
F2 synform, antiform; with plunge
Overturned F2 antiform
Highway
41

**Figure 19.82.** Sketch map showing structural trends within the Mazinaw Terrane and adjacent terranes.

**Table 19.17.** Comparative geologic histories of the Elzevir, Mazinaw and Sharbot Lake terranes

| **Elzevir Terrane** | **Mazinaw Terrane** | **Sharbot Lake Terrane** |
|---|---|---|
| | | <1050 million-year-old mylonitization along Robertson Lake mylonite zone |
| 1090 million-year-old Skootamatta Suite intrusions | | 1090 million-year-old Skootamatta Suite intrusions? |
| Ottawan orogenic effects minimal? | regional metamorphism to upper amphibolite facies associated with Ottawan Orogeny | Ottawan orogenic effect minimal |
| | uplift, rifting and sedimentation ca. 1160–1150 Ma (Flinton Group deposition) | |
| regional metamorphism coincident with Elzevirian Orogeny | regional metamorphism, thrusting, shearing associated with Elzevirian Orogeny | regional metamorphism coincident with Elzevirian Orogeny |
| 1250 million-year-old granite and gabbro-diorite plutonism | 1250 million-year-old? granite plutonism | 1250 million-year-old? gabbro-diorite and granite-granodiorite plutonism |
| 1260–1250 million-year-old bimodal tholeiitic volcanism, volcaniclastic and carbonate sedimentation (Hermon and Mayo groups, Grimsthorpe Group) | 1260–1250 million-year-old? calc-alkalic volcanism, volcaniclastic and carbonate sedimentation, (Mazinaw Group) | 1260–1250 million-year-old? tholeiitic volcanism, volcaniclastic sedimentation and carbonate sedimentation (Sharbot Lake Group) |
| 1270 million-year-old tonalite plutonism | basement of ca. 1270 million-year-old tonalite and gabbro | ? ? |
| 1290–1270 million-year-old gabbro plutonism | | |
| >1290 million-year-old gabbro and tholeiitic basalt magmatism (Canniff and Basal groups) | | |

represent Bishop Corners, and Myer Cave formation metapelites.

Figure 19.82 shows the distribution of Flinton Group strata within the Mazinaw Terrane. The Flinton Group locally lies unconformably on rocks of the Mazinaw ("Grenville") Group (*see* Figure 19.82 for localities). As noted above, the Flinton-Mazinaw group unconformity is locally preserved. In addition, mafic dikes common in the Mazinaw Group do not cut the Flinton Group, and other than a few late syenite pegmatite dikes, no intrusions cut the Flinton Group (Moore and Thompson 1980). These observations are consistent with the unconformable nature of the Flinton Group. Detailed studies of the conglomerates (Walton et al. 1964; van de Kamp 1971; Psutka 1976) also suggested derivation from underlying plutonic rocks.

Moore and Thompson (1972, 1980), Harnois (1987) and Harnois and Moore (1988) describe a regolith locally preserved beneath the Flinton Group. The regolith is preserved as a staurolite-garnet schist, termed the Ore Chimney Formation (Moore and Thompson 1980), which is part of the Mazinaw Group. The extent of the regolith is unclear, as some workers regard the Ore Chimney Formation as originating due to metasomatism in a shear zone (e.g., Dillon 1985; Laing et al. 1987).

Rocks directly correlative with the Flinton Group are not present within the Elzevir or Sharbot Lake terranes. This restriction of the Flinton Group to the Mazinaw Terrane (Elzevir Terrane *of* Moore 1982) is anomalous, because it means that this unconformable sequence is only preserved in the most structurally complex, highest grade part of Moore's (1982) Elzevir Terrane. If the Mazinaw Terrane has had a unique and separate geologic history compared to the Elzevir and Sharbot Lake terranes, then the restriction of the Flinton Group to the Mazinaw Terrane is easily explained.

The Flinton Group shows evidence of being derived from both local and more distal sources. South of Sand Lake and southwest of Shawenegog Lake, granite boulders and feldspathic litharenites rim the Norway Lake Granite, the probable source of the boulders (Smith 1958; Smith et al. 1969; Easton and Ford 1991). Easton and Ford (1991) found that basal conglomerates were more extensive than previously recognized, and contain interbedded pelitic and

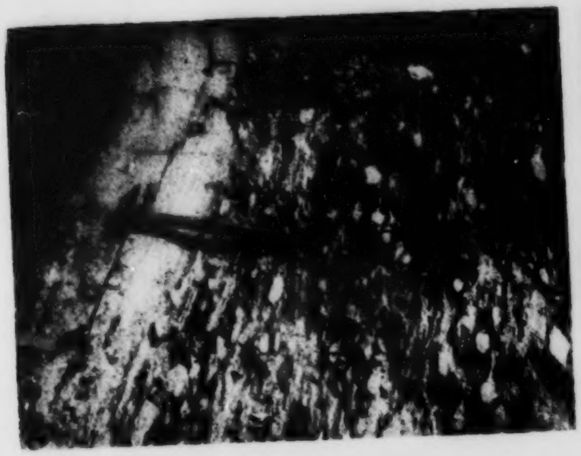


**Figure 19.83a.** Quartz arenite (left) and quartz arenite cobble conglomerate (right) of the Bishop Corners Formation, Flinton Group, Highway 41 north of Northbrook. Hammer handle is 30 cm long.

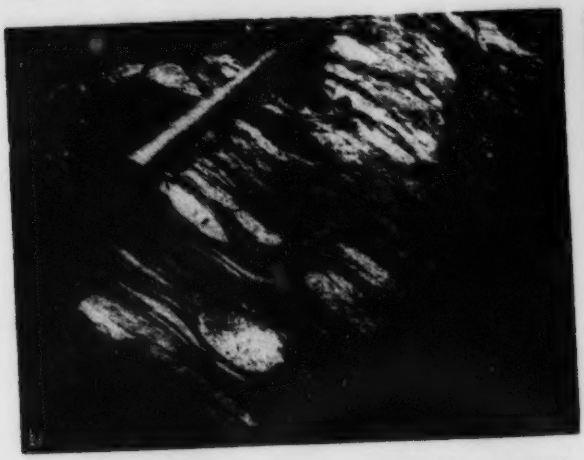


**Figure 19.83b.** Stretched cobble conglomerate within the Lessard Formation, Highway 41, north of Kaladar. Hammer handle is 30 cm long.

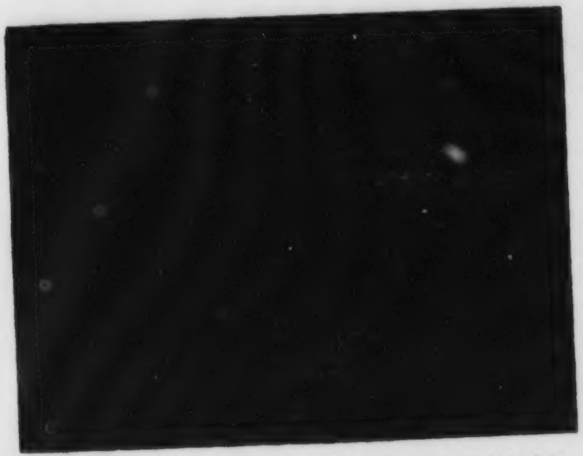


**Figure 19.83c.** Thinly laminated sulphide-bearing black schist of the lower Myer Cave Formation, Flinton Group, Highway 506 east of Marble Lake. Field of view is 20 cm across.

**Figure 19.83d.** Poorly laminated sulphide-bearing black schist of the upper Myer Cave Formation, Flinton Group, Highway 506 west of Fernleigh. Hammer handle is 30 cm long.

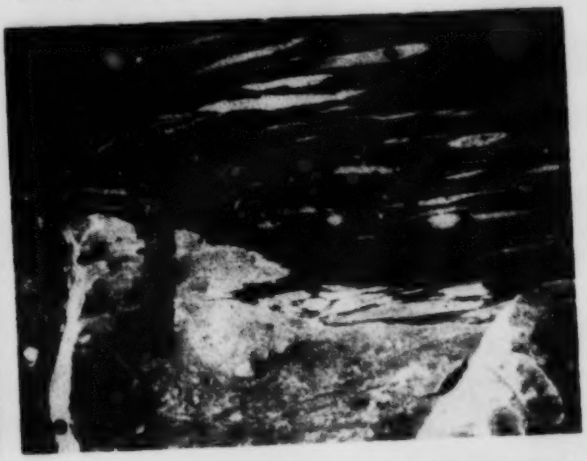


**Figure 19.83e.** Carbonate breccia unit with black shale matrix, Myer Cave Formation, Flinton Group, Highway 506 east of Marble Lake. Hammer handle is 30 cm long.

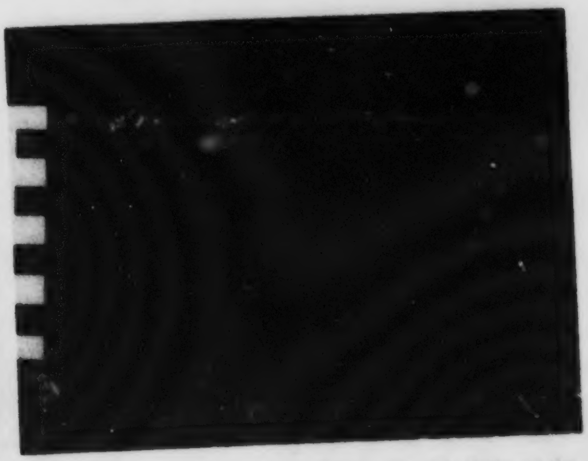


**Figure 19.83f.** Thinly bedded calcareous silts and shales of the Fernleigh Formation, Flinton Group, with microcline porphyroblasts. Scale bar is 9 cm long.

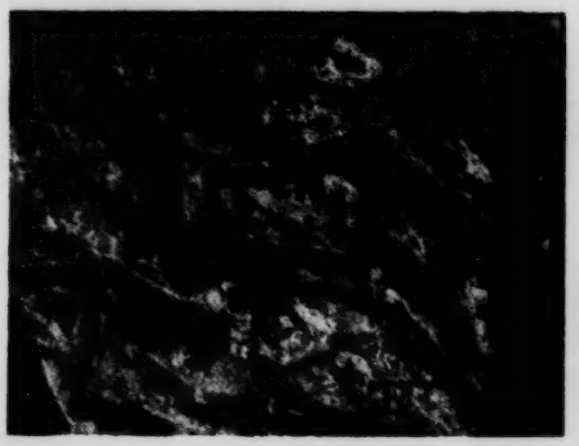


**Figure 19.84.** Kyanite porphyroblasts within biotite-kyanite-staurolite schists of the Bishop Corners Formation near Fernleigh. Hammer head is 17 cm across.

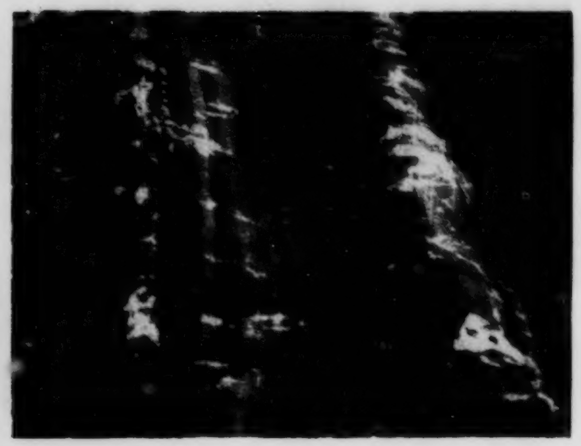


**Figure 19.85a.** Thin- to medium-bedded calcitic marbles of the Myer Cave Formation, Flinton Group, Highway 506 east of Fernleigh. Hammer handle is 30 cm long.

semi-pelitic schists, including distinctive biotite-kyanite schists. The conglomerates are in unconformable contact with the granites, marbles and amphibolites in the area, thus, the conglomerates are correlated with the Flinton Group. Significantly, these Flinton Group occurrences are not in the typical synclinal setting that Moore and Thompson (1972, 1980) document for the Flinton Group, and the Flinton Group thus may not be as infolded with Mazinaw Group strata as previously believed (cf., Rivers 1976; Chappell 1978). Moore and Thompson (1980) also note conglomerates containing tonalite and granite clasts adjacent to the Northbrook Tonalite and Addington Granite, near Flinton and Kaladar respectively, are not associated with major synclines. Moore and Thompson (1972, 1980) also describe major facies changes in Flinton Group rocks suggesting significant local source controls. Kinsman and Parrish (1990) suggest a more distal source region, as 1157 million-year-old zircons in Bishop Corners Formation metamorphosed quartz arenites are similar in age to Frontenac Terrane plutons. Frontenac Terrane metamorphosed quartz arenites may also be the source of quartz arenite cobbles within the Flinton Group. Alternatively, the Abinger Granite (Rb-Sr age of 1185±25 Ma, Bell and Blenkinsop 1980) and the Norway Lake Granite, which locally shed detritus into the Flinton Group, might be the source of the 1157 million-year-old zircon population reported by Kinsman and Parrish (1990).

Moore and Thompson (1972, 1980) suggested a fluvial environment for the coarse clastic units of the Flinton Group and a shallow-marine setting for the pelitic and carbonate units, both being deposited in an extensional environment. Easton and Ford (1991) postulate a fluvial-lacustrine environment for the Flinton Group in the Mazinaw Terrane, with deposition controlled by rift valleys resulting in longitudinally oriented alluvial-fan and braided stream trunk systems (Figure 19.86). This model explains the preservation of the Flinton Group in synformal, linear belts; localized faulting present along the Flinton unconformity; and localized mineralization. Alternating pelite-carbonate beds within the Fernleigh Formation may reflect annual variation in sediment supply into rift valley lakes, and the ubiquitous and abundant presence of tourmaline throughout the Flinton Group (e.g., Thompson 1972; Easton 1988a) would be expected in an arid depositional environment. The Myer Cave Formation megabreccia unit probably formed by collapse of a cliff into a lake, allowing for the incorporation of Myer Cave Formation and Mazinaw Group blocks in a pelitic matrix. The depositional setting outlined above is similar to that described by Lefebvre (1989) for the Zambian copper deposits, as are the metal associations (gold, copper, lead, zinc and arsenic). The metallogenic implications of this model are outlined *in* Easton and Fyon (this volume).

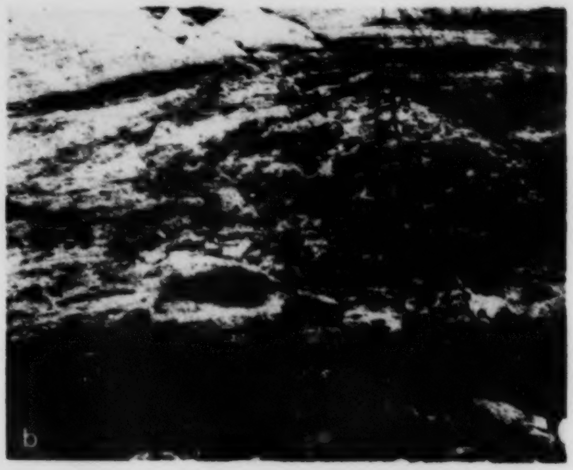


**Figure 19.85b.** Domal stromatolite (right) within dolomitic marbles of the Flinton Group, Clare River area, Highway 41 south of Kaladar. Field of view is 20 cm across.

## GEOCHEMISTRY OF METAVOLCANIC ROCKS

Two chemical suites are represented in the Mazinaw Terrane, a tholeiitic suite and a calc-alkalic suite, studied in the Harlowe–Marble Lake area by Sethuraman (1970), Sethuraman and Moore (1973), Condie and Moore (1977), Ayer (1979), Moore and Morton (1986) and Harnois (1987). Chappell (1978) and Bright (1986a) discusses the chemistry of metavolcanic rocks within the Clare River synform in the southern part of the terrane.

The tholeiitic succession straddles the Grimsthorpe Domain–Mazinaw Terrane boundary, and thus, some of the analyzed tholeiitic rocks are from the Grimsthorpe Domain. The calc-alkalic succession, however, lies entirely within the Mazinaw Terrane. Thus, continuity of the 2 successions is debatable; more likely, the 2 sequences are tectonically juxtaposed (Easton and Ford 1991). Neither metavolcanic sequence has been dated.

Condie and Moore (1977) and Sethuraman and Moore (1973) classify the tholeiitic sequence as low-potassium tholeiites, similar to modern arc and rise tholeiites and Archean low-potassium tholeiites (*see* Figure 19.62). Little chemical variation is observed in the tholeiitic sequence. The tholeiites are nepheline normative and contain chlorite (Sethuraman and Moore 1973). Both features could be secondary (Moore and Morton 1986). Mafic volcanic rocks near Flinton are also tholeiitic (Wolff 1982a), in agreement with the results of Sethuraman and Moore (1973). The rocks near Flinton, however, are most likely part of the Grimsthorpe Domain.

The calc-alkalic sequence consists of andesites, dacites and minor rhyolite and are similar to modern calc-alkalic suites (Condie and Moore 1977; *see* Figure 19.62). Fragmental rocks are common in the intermediate and felsic part of the sequence. Dacite dikes cut the tholeiitic sequence and the Elzevir Tonalite and have been interpreted as feeders to the calc-alkalic volcanic succession (Moore and Morton 1986; Lumbers et al. 1990). Possibly correlative dacite dikes, adjacent to the Elzevir Tonalite, have been dated at $1229^{+11}_{-4}$ Ma (Connelly et al. 1987).

Ayer (1979), in the southern Mazinaw Lake–Pringle Lake area, reported a tholeiitic basalt-andesite-rhyolite complex, which he considered to be a volcanic centre. Rocks in this area are generally east-trending and are structurally discordant with the north- to northeast-trending calc-alkalic metavolcanic sequence to the south (Moore and Morton 1986). Easton and Ford (1991) remapped much of the area and reported the presence of few recognizable volcanic textures within the sequence. Easton and Ford (1991) described these rocks as laminated, quartzofeldspathic gneisses that may in fact be recrystallized mylonites (*see also* Rivers 1976). Some rocks adjacent to the Norway Lake Granite were reinterpreted by Easton and Ford as flattened conglomerates of the Flinton Group. Thus, the extent of the Mazinaw Lake–Pringle Lake volcanic complex may be less than originally indicated by Ayer (1979) and Moore and Morton (1986).

Bright (1986a) reports basaltic and dacitic-rhyolite metavolcanic rocks in the Clare River synform (Shovel Lake Formation *of* Chappell 1978) and in the Salmon River area. Metavolcanic rocks from both occurrences are bimodal, consisting of tholeiitic basalts and tholeiitic to slightly calc-alkalic dacite and rhyolite.

In the western half of the Darling area, mafic, intermediate and felsic metavolcanic rocks are metamorphosed to upper amphibolite facies and are intensely deformed by the adjacent Robertson Lake mylonite zone and a high-strain zone along the eastern margin of the Bartraw dome (Easton 1988a). Limited geochemistry on this sequence suggests a calc-alkalic affinity, however, abundant carbonate in these rocks makes interpretation problematic (Easton 1988a). These rocks continue south into the Lavant area (Pauk 1989b), but no geochemistry is available from that area.

In summary, the Mazinaw Terrane volcanic sequences are largely calc-alkalic, unlike the dominantly tholeiitic Elzevir Terrane metavolcanic rocks.

## PLUTONISM

It has been assumed previously that the plutonic rocks of the Mazinaw Terrane, such as the Northbrook/Cross Lake Tonalite and the Abinger Granite, were equivalent to the Elzevir and Weslemkoon tonalites and the Methuen Suite in terms of age, composition and origin. However, apart from the Addington Granite which has been dated at $1245^{+15}_{-10}$ Ma (van Breemen and Davidson 1988b), no U-Pb zircon ages are available to confirm this assumption. Skootamatta Suite plutons, which are common in the Elzevir and western Frontenac terranes, are notably absent from the Mazinaw Terrane. Unlike the Elzevirian domains, the Mazinaw Terrane is characterized by elongate, northeast-trending lenticular plutonic patterns, exemplified by the Addington Granite and the Northbrook Tonalite (*see* Figure 19.82).

In the central part of the terrane, a repetitious pattern of elongate tonalite masses separated by zones of layered amphibolite, para-amphibolite and siliceous clastic metasedimentary rocks, exemplified by the Northbrook and Cross Lake tonalites, is also suggestive of a tectonically influenced shape. The tonalite plutons are associated with the metavolcanic rocks of the terrane. The Mazinaw Terrane tonalites differ from Grimsthorpe Domain tonalites in being more granodioritic in composition (e.g., Wolff 1982a; Kamilli 1974), being more deformed and being more migmatitic. Most of these differences could be attributed to the higher grade metamorphic conditions seen in the Mazinaw Terrane. The geochemistry of the tonalites has been studied by Lumbers et al. (1990). Somers (1984) studied the White Lake Tonalite located in the extreme northeast corner of the terrane.

The best evidence for tectonism being responsible for pluton form is found near Coxvale, where the northern and southern lobes of the Cross Lake Tonalite are separated by a high-strain zone developed in both tonalite and supracrustal rocks, and where an increasing strain gradient can be observed in the tonalite as the contact is neared. The form of the Cross Lake Tonalite is also suggestive of a folded thrust contact, rather than an unmodified intrusive contact. Lumbers et al. (1990) report an age of *circa* 1270 Ma from the Cross Lake Tonalite, suggesting that the tonalites may have served as basement to the metavolcanic sequence.

The granitic bodies within the Mazinaw Terrane are characterized by a fine- to medium-grain size and considerable compositional heterogeneity ranging from granodiorite and quartz monzonite to syenogranite. Monzogranites and monzonites predominate. Muscovite is common in all phases of these leucocratic granitic rocks, which are typically weakly foliated to gneissic. Bright (1986a) presents geochemical data on the Mellon Lake Complex. Moore (1967), Wolff (1982a), Carter (1981) and Easton (1988a) present geochemical and modal data for the Addington Granite, and Lumbers et al. (1990) give an average analysis for the Abinger Granite. Lumbers et al. (1990) include both the Addington and Abinger granites in the alaskite (Methuen) suite, even though they are considerably more heterogeneous and include a greater proportion of metasedimentary xenoliths than Methuen Suite granites in the Elzevir Terrane.

In the northern part of the terrane, the Abinger and Norway Lake granites have tectonized northern margins and are separated by narrow belts of supracrustal rocks. The shape of the granitic plutons, in part, is largely a reflection of tectonism. This is also consistent with gravity models across the Mazinaw Terrane (Real and Thomas 1987), which indicate that the plutons are extremely thin (approximately 1 km thick).

The granitic plutons are associated with the dominantly metasedimentary belts within the terrane. The granitic bodies commonly contain inclusions and inclusion trains of country rocks. Given the lithologic variation between phases of the granite separated by inclusion trains, the granite bodies may have been emplaced as a series of sills. Wolff (1982a) reached the same conclusion regarding the emplacement of the Addington Granite. The Abinger Granite intrudes tonalite intrusion breccias along Highway 41, indicating that this body is younger than the tonalites. The granite is more deformed than the intrusion breccia it intrudes, perhaps due to syndeformational emplacement.

The Mellon Lake Complex (Bright 1986a; Sheffield intrusion *of* Wolff 1982a) has also intruded an older, migmatitic, nebulitic tonalite, and both Bright (1986a) and Wolff (1982a) have suggested that it may, in part, have served as basement to the supracrustal sequence in the Clare River area. The Bartraw dome (Karboski 1980; Easton 1988a) in the northeast of the Mazinaw Terrane is cored by migmatized tonalite and rimmed by a medium-grained, alaskitic granite. On structural grounds, Karboski (1980) suggested that the Bartraw dome may have also served as a basement complex.

The Elbow Lake meta-anorthosite (Lumbers 1982; Lumbers et al. 1990) on the southeast shore of Norcan Lake bears an unknown relationship to the anorthosite suite rocks in the Central Metasedimentary Belt. Anorthosite slivers also occur south of this body in the tectonized margin of the Bartraw dome (Easton 1988a).

Rb-Sr whole-rock ages on the Abinger and the Mellon Lake bodies are 1185±25 and 1120±90 Ma respectively (Bell and Blenkinsop 1980), although the preliminary age on the Mellon Lake Complex was 1250±130 Ma (Bell and Blenkinsop 1979). These ages could have been reset during regional metamorphism, as the Rb-Sr age on the Addington Granite 1035±42 (Krogh and Hurley 1968) is some 200 million years younger than the U-Pb zircon age on the

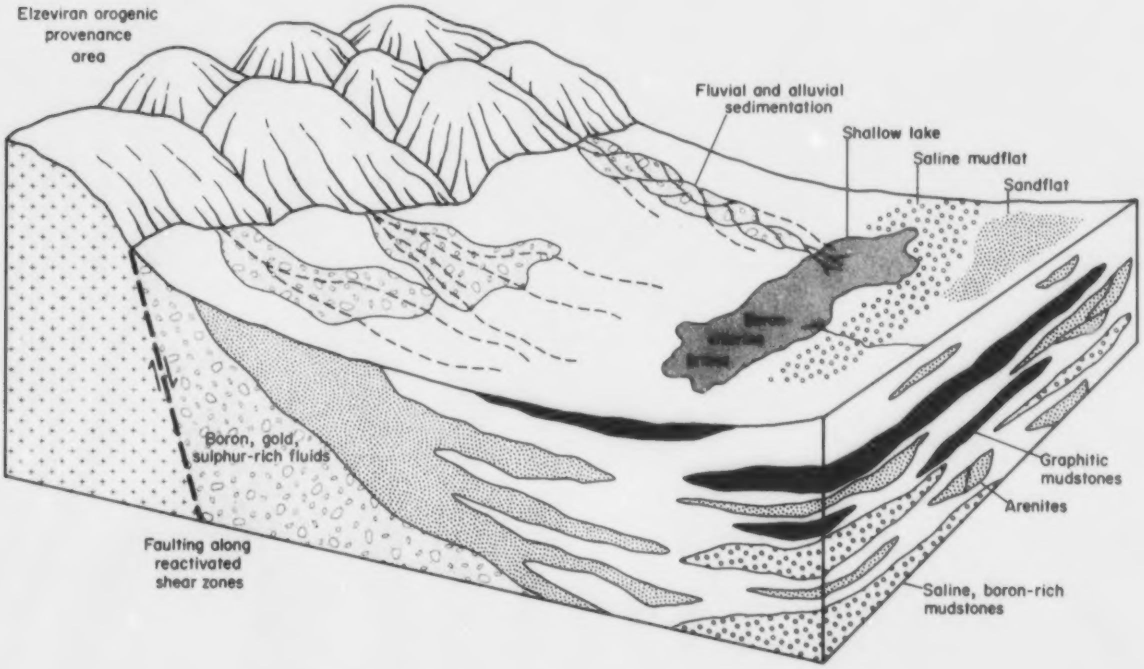

Elzeviran orogenic provenance area
Fluvial and alluvial sedimentation
Shallow lake
Saline mudflat
Sandflat
Boron, gold, sulphur-rich fluids
Faulting along reactivated shear zones
Graphitic mudstones
Arenites
Saline, boron-rich mudstones

**Figure 19.86.** Sketch showing a possible depositional model for the Flinton Group.

same body. However, the presence of 1160 to 1150 million-year-old zircons in the Flinton Group, which seems to be largely of local provenance, may indicate that the Rb-Sr ages approximate the age of emplacement of the granitic component of these bodies. If so, then 2 ages of granite magmatism may be present in the area: a *circa* 1240 million-year-old magmatism comparable in age with the Methuen Suite (e.g., Addington Granite), and a *circa* 1160 million-year-old magmatism similar in age to Frontenac Terrane magmatism but of alaskite suite affinity (e.g., Abinger and Norway Lake granites).

## STRUCTURAL HISTORY

Several structural studies have been conducted in the western and central part of the Mazinaw Terrane, most notably Venkitasubramanyan (1969), Thompson (1972), Rivers (1976), Chappell (1978) and Wolff (1982a). In general, 3 main periods of deformation resulting in folding have been recognized. These 3 periods of deformation are generally considered to have affected all supracrustal rocks in the Mazinaw Terrane, including the Flinton Group. $D_1$ produced $F_1$, northeast-trending isoclinal folds. An earlier deformation event ($D_0$) is recorded only in fragments within tonalite intrusion breccias in the area (Moore and Thompson 1980). $D_2$ also produced northeast-trending folds ($F_2$), generally coaxial to the $F_1$ folds; tighter structures; and local overturning of stratigraphy. The $F_2$ folds are readily evident on geologic maps of the area and include the Bishops Corners, Plevna and Ompah synclines.

Chappell (1978) suggested that thrusting in the Clare River area occurred late in $D_1$ or was post-$D_1$, and that the Clare River synform was a nappe. The décollement surface was located in a narrow metasediment band adjacent to the Clare River Group–Addington Granite contact. The décollement surface was folded during $D_2$.

Easton and Ford (1991) conclude that thrusting in the Mazinaw Lake area occurred during, or prior to, $D_2$ and $F_2$. The Norway Lake Granite is bounded by a shear zone to the south and north and may be a nappe. This shear zone also truncates the north limb of the Ompah syncline and the south limb of the Plevna syncline (Easton and Ford 1991). Chappell (1978) and Rivers (1976) both noted that metamorphic minerals overgrow the $D_1$-$S_1$ fabric in the Clare River and Fernleigh–Ompah areas, suggesting that thrusting may have occurred prior to peak metamorphism. This suggests that heating of crust was relatively slow, and is consistent with the observation that metamorphism and plutonism are not coincident.

The deformation history of the Flinton Group is more difficult to unravel. Rivers (1976) and Chappell (1978) both questioned the relation of the Flinton Group and the $F_2$ synclines, and Easton and Ford (1991) confirm that the Fernleigh belt is a northeast-dipping, southeast-facing homocline as suggested by Rivers (1976). Thus, the localization of the Flinton Group in synformal troughs may be an artifact of deposition; consequently, the synclines post-ulated by Moore and Morton (1986) do not exist. Depending on the timing of deformation, the Flinton Group may have been largely unaffected by the $F_2$-$D_2$ event. It was affected mainly by $D_3$, a period of coaxial, northeast-trending open folding and cleavage development and by $D_4$, a period of broad, northwest-trending open folds ($F_4$).

Connelly (1985, 1986) and Connelly et al. (1987) used structural evidence and geochronologic data in the Flinton area to suggest that the main fabric forming event in the region affected the Elzevir Tonalite, "Tudor metavolcanics" and the Flinton Group, and suggested that the Flinton Group was tectonically juxtaposed against the Elzevir Tonalite during the final stages of its crystallization. Certainly in the area studied, the Flinton Group is in tectonic contact with mafic metavolcanic rocks and the Elzevir Tonalite (Easton, unpublished data). If the major Grenville metamorphic event occurred at *circa* 1230 to 1180 Ma, then the major deformation fabric would be in all units, as noted by Connelly (1986), negating the argument that the Flinton Group be the same age as the Mazinaw Group (Connelly 1986).

The structural history of the Mazinaw Terrane is similar to that outlined by deLorraine and Carl (1986) for the Frontenac Terrane. Namely, a $D_1$ event resulting in northeast isoclines ($F_1$) and migmatization ($M_1$); $D_2$ thrusting and broad, northeast-trending folding ($F_2$); $D_3$ north-northeast-trending tight to isoclinal folds ($F_3$) resulting in local overturning and finally; $D_4$, resulting in gentle, northwest-trending folds ($F_4$). Wynne-Edwards (1967a, 1967b) outlined a similar structural history, although the more intense metamorphism in the Frontenac Terrane in Ontario apparently completely transposed the $F_1$ and $F_2$ events.

## METAMORPHIC HISTORY

The metamorphic history of the Mazinaw Terrane is complex. Two, perhaps 3, regional metamorphic episodes affected these rocks; 2 prior to Flinton deposition. The first metamorphic event affected the metavolcanic and metasedimentary xenoliths preserved within tonalite intrusion breccias in the area. This may have been a contact metamorphic event. At least 1 regional metamorphic event preceded Flinton Group deposition. It was accompanied by high metamorphic temperatures and intense deformation including thrusting and folding. This event caused migmatization in and around the Abinger Granite and probably attained middle to upper amphibolite-facies conditions across most of the Mazinaw Terrane. During this episode, thrusting of metaplutonic masses into their current lenticular, northeast-trending pattern occurred and shear zones developed along the northern margins of most metaplutonic masses in the terrane. In addition, Mazinaw Group limestones were locally metamorphosed to dolomite marbles, which were subsequently eroded to form dolomite clasts present in Flinton Group conglomerates. This metamorphic event may have occurred in the period 1230 to 1180 Ma and may have been even older, perhaps coincident with Elzevir Terrane magmatism (*circa* 1240 Ma).

Much of the structural framework of the area was already established prior to Flinton Group deposition, as all shear zones have been overprinted by the latest regional metamorphism, which affects the Flinton Group. This latest regional metamorphism was one of lengthy heating with only limited deformation, mainly regional folding and cleavage development, and was followed by a period of slow cooling. The result of this lengthy heating episode has been extensive recrystallization, resulting in a great textural similarity, and lack of primary structures, in all rocks in the domain. A variety of porphyroblasts (*see* Figure 19.84), including biotite, plagioclase, magnetite, garnet, sillimanite, kyanite and staurolite, developed in rocks of appropriate bulk composition and overprint the structural fabric in the rocks. Conditions were upper greenschist to middle amphibolite facies (cf., Chappell 1978). In the Clare River area, Bright (1986a) and Chappell (1978) both locally document a lower metamorphic grade assemblage overprinting a higher grade assemblage.

## BASEMENT

Two basement-cover relationships may be observed in the Mazinaw Terrane: 1) the Flinton Group metasedimentary rocks were deposited on a deformed and metamorphosed Grenville supracrustal and plutonic basement; and 2) tonalitic rocks may have served as basement to the Mazinaw Group. In the latter case, the Abinger Granite intrudes ultramafic and metagabbroic, and metatonalitic, rocks similar to those found in the Grimsthorpe area to the west (Easton and Ford 1991). In addition, slivers of ultramafic and metagabbroic rocks are found in the belts of layered quartzofeldspathic gneisses separating supracrustal-granite packages in the northern Mazinaw Terrane (Easton and Ford 1991). These mafic and ultramafic rocks are similar to rocks in the Grimsthorpe Domain, greater than *circa* 1270 million years (Lumbers et al. 1990) in age. This relationship may also imply that the Northbrook, Cross Lake and other tonalite bodies in the Mazinaw Terrane are similar in age to the Elzevir and Weslemkoon tonalites. It is critical to the tectonic history of the area to know if the volcanic rocks in the Harlowe–Marble Lake area are similar in age to those in the Belmont Domain (ca. 1250 Ma and hence younger than the tonalites) or are indeed older than the tonalites. To the south, Bright (1986a) has suggested that the Mellon Lake Complex of tonalite and younger granitic phases may also served as basement to supracrustal rocks in the Clare River synform.

## MINERAL DEPOSITS

Many mineral deposit types are absent or rare in the Mazinaw Terrane, most notably contact-metasomatic iron deposits and copper-nickel deposits, both of which are associated with Lavant Suite gabbro-diorite plutons, which are uncommon in the Mazinaw Terrane. The Radenhurst–Caldwell deposit (Carter et al. 1980; Easton 1988a) consists of massive magnetite hosted in mafic metavolcanic and synvolcanic metagabbroic rocks and is probably the best described deposit of this type. It contains elevated light REE contents (1 to 2% LREE, Easton 1988a), an atypical feature of Central Metasedimentary Belt magnetite deposits.

Two broad groupings of metallic mineral deposits are present in the terrane. The first consists of sulphide mineralization (pyrite, pyrrhotite, chalcopyrite, arsenopyrite, magnetite and bornite) associated with metavolcanic rocks, generally as disseminations or within interflow metasedimentary rocks (e.g., Blithfield pyrite deposit, Carter et al. 1980; Easton 1988a). The second consists of quartz-carbonate-sulphide vein and sulphide mineralization found in close proximity to the Flinton Group unconformity and within the Flinton Group itself. Most significant mineral deposits in the terrane are of the second type.

Vein mineralization associated with the Flinton Group unconformity varies with host rock composition. In mafic metavolcanic host rocks, gold-arsenopyrite-pyrite mineralization is typically associated with quartz ± carbonate veins (e.g., Addington Mine, Papertzian 1984, Dillion 1985, Laing et al. 1987, Harnois 1987, Harnois and Moore 1989; Ore Chimney Mine, Meen 1944, Papertzian 1984, Moore and Morton 1986; O'Donnell occurrences, Moore and Morton 1986). Metawackes also host similar gold-bearing vein systems (e.g., Boerth and Webber properties and the James Mine, Smith 1958, Pauk 1987). Carbonate rocks also host gold-pyrite-sphalerite-galena-bearing, dolomite-quartz vein systems near the unconformity (e.g., Cook, Helena, Stead, Camgar, Dome and Pay Rock gold occurrences, Papertzian 1984, Moore and Morton 1986).

Easton and Ford (1991) noted that most gold occurrences in the northern Mazinaw Terrane are associated with an older shear zone near the Flinton Group unconformity. This older shear zone may have been a critical factor in siting mineralization and influencing subsequent Flinton Group deposition. Although most of these deposits are located near the unconformity with the Flinton Group, the variability in mineralization types and host-rock and vein compositions, suggests that the specific genetic process varied considerably between deposits. Weathering near the unconformity surface may have locally concentrated metals which were latter mobilized into veins during metamorphism. Faulting and fluid movement associated with the extensional environment in which the Flinton Group was deposited may have been an important factor in mineralization, especially as nondetrital tourmaline is closely associated with Flinton Group metasedimentary rocks and many of these deposits. In addition, fluids associated with the Flinton Group may also have aided in metallogenesis, with the unconformity serving as a fluid channel.

Mineralization is found in 4 settings within the Flinton Group itself. 1) In the Marble Lake area, black pyritic-graphitic schists of the Myer Cave, and, locally, the Fernleigh formation contain stratiform pyrite-chalcopyrite-sphalerite mineralization. In the Clare River synform, similar mineralization has been described by Bright (1986a) from the Kaladar Mines Ltd. property and the Amiko mica property. 2) In the Clare River synform, pyrite-pyrrhotite deposits are closely associated with basal muscovite-schists of the Flinton Group (Bright 1986a). Examples include the Canada and Hungerford mines near Sulphide (Bright

1986a) and the Ontario Sulphur and Donahue Creek deposits (Carter 1984). 3) Bright (1986a) reports stratabound zinc mineralization with minor pyrite and chalcopyrite in dolomitic marbles in the Clare River area, possibly equivalent to the Myer Cave Formation (e.g., Spry Zinc occurrence, Carter 1984; possibly the Glenshire property, Carter 1984, Bright 1986a). 4) Best known in both areas are crosscutting sulphide-quartz veins in the dolomite horizon of the Myer Cave Formation containing pyrite-chalcopyrite-sphalerite-galena mineralization. Examples include the Barrie Syndicate and Mazinaw Base Metals occurrences (Moore and Morton 1986) and the Ganda Silver mines occurrence (Pauk 1987). This mineralization is probably the result of remobilization of Kuperfiescher-type, copper-zinc mineralization related to the original deposition of the Flinton Group and represented by types 1 and 3. The Zambian copper belt (e.g., Lefevbre 1989) is similar in many respects to the postulated environment for Flinton Group deposition and mineralization both within, and adjacent to, the Flinton Group. The low sulphur isotope content of the Spry zinc deposit (2.6‰, Sangster 1991) is consist with it being derived from basinal brines that flowed through adjacent metavolcanic strata (Sangster 1991). It is noteworthy that lead is more abundant in Mazinaw Terrane sulphide occurrences than elsewhere in the Central Metasedimentary Belt. This may be a reflection of the fact that supracrustal rocks of the Mazinaw Terrane were deposited on a basement complex of tonalite and gabbroic rocks. Although this basement is only slightly older than the overlying supracrustal rocks, it may have served as a source of lead.

The Clyde Forks copper-antimony-silver-mercury-barite occurrence (Nikols 1972; Carter et al. 1980; Carter 1984) represents a unique deposit type in the Central Metasedimentary Belt. It is a conformable deposit hosted in dolomitic marbles and quartzofeldspathic metasedimentary rocks of the Mazinaw Group, close to the trace of the Clyde River fault. Nikols (1972) proposed that mineralization resulted from hydrothermal springs discharging into a carbonate basin. Flinton Group rocks are present about 1 km east of the deposit, and it is not known if hydrothermal activity related to Flinton Group deposition played a role in the genesis of this deposit (in favourable host rocks).

Uranium- and thorium-bearing pegmatite veins occur in 3 concentrations in the easternmost Mazinaw Terrane: 1) the Cross Lake area (Ford and Charbonneau 1979; Ford 1982), 2) near Barnett Chute (Easton 1988a) and 3) the Salmon River area (Bright 1986a). Compared to Bancroft Terrane pegmatites, the Mazinaw Terrane pegmatites have distinctly lower regional equivalent thorium values (less than 5 ppm; Ford 1982). The concentration of pegmatites in these areas may reflect partial melting of metaplutonic rocks (and possibly adjacent paragneisses, cf., Ford 1982) related to upper amphibolite metamorphism (all are located in large plutonic complexes in areas of peak metamorphic conditions). The lower thorium content may reflect their derivation from tonalitic to granodioritic metaplutonic rocks (Mellon Lake Complex, Cross Lake Tonalite, Bartraw dome). As in the case of other Grenville uranium vein deposits, gamma-ray spectrometry has identified all potential targets in the Central Metasedimentary Belt, and the most favourable deposits are those found in pegmatites hosted in adjacent metavolcanic rocks (cf., Bright 1986a; Pauk 1987). Pauk (1987) describes the exploration history of the higher grade Cross Lake occurrences. Pauk (1987) reports higher uranium values at surface then at depth, probably due to supergene enrichment during weathering. Locally, molybdenum mineralization is present in association with the pegmatites, most commonly in the less radioactive pegmatites.

## SHARBOT LAKE TERRANE

The Sharbot Lake Terrane consists of mafic metavolcanic and carbonate rocks intruded by gabbros, granodiorite to granite plutons and the Dalhousie Lake amphibolite complex (Pauk 1989b). The Hinchinbrooke Granodiorite Gneiss south of the Sharbot Lake Terrane may in fact be part of the Mazinaw Terrane. The terrane has been completely mapped at 1:50 000 and 1:31 680 scale (Dugas 1950; Wilson and Dugas 1961; Reinhardt et al. 1973; Hill 1974; Wolff 1982b, 1985; Easton 1988a, 1990d; Pauk 1989a, 1989b). Metavolcanic rocks and volcaniclastic metasedimentary rocks, metawackes and semipelites underlie roughly 10 and 15% of the Sharbot Lake Terrane, respectively. Metaplutonic rocks underlie about 30% of the terrane, with marbles comprising the remaining 45%. Moore (1982) defined the Sharbot Lake Terrane as a predominantly carbonate basin. The Sharbot Lake Terrane is lithologically similar to parts of the Elzevir Terrane, and the 2 are probably closely related. No geochronology is available from the Sharbot Lake Terrane.

### STRATIGRAPHIC RELATIONS

#### Sharbot Lake Group

Supracrustal rocks in the Sharbot Lake Terrane are assigned to the Sharbot Lake Group (new term) (*see* Figure 19.54 and Table 19.7). No formations are here proposed for the Sharbot Lake Group.

Metavolcanic rocks are best exposed south of Darling Lake and consist mainly of aphanitic rocks, commonly cut by carbonate veins. Locally, thin-selvaged pillows are preserved, as are beds of tuff breccia, lapilli tuff and thinly to thickly bedded metatuffs. Thickness of the volcanic sequence is over 300 m (Easton 1988a).

In the Darling area, the metavolcanic succession is overlain by roughly 100 to 300 m of fine-grained, thinly bedded, silty to sandy clastic metasedimentary rocks, black shales, felsic tuffs and cherty beds of mainly volcanic exhalative provenance, which are locally pyritiferous and grade into sulphide facies ironstones (Easton 1988a). Farther south, between Sharbot Lake and Carleton Place, this metasedimentary sequence is considerably thicker and includes para-amphibolite, probably derived from mafic volcaniclastic metasedimentary rocks (Reinhardt et al. 1973; Wolff 1982b, 1985; Easton and McMaster 1987).

These clastic metasedimentary rocks are overlain by calcitic and dolomitic marbles which show considerable textural variation. Dolomitic marble, locally containing algal-laminite stromatolites, predominate in areas adjacent to the metavolcanic rocks (Easton 1988a), with thinly bedded, graded, calcite turbidites and their high grade equivalents forming the bulk of the carbonate belt in the terrane. In general, marbles in the eastern part of the terrane are relatively free of impurities.

## GEOCHEMISTRY OF THE METAVOLCANIC ROCKS

In the northern part of the Sharbot Lake Terrane, Easton (1988a) documents the tholeiitic affinity of the basaltic sequence. Carbonate is commonly disseminated throughout the basaltic sequence, making geochemical characterization difficult. To the south, Wolff (1982b) reports the mafic rocks are tholeiitic basalts. A few felsic rocks are present, but there are too few analyses to reliably classify them.

## PLUTONISM

The Sharbot Lake Group is intruded by fine- to coarse-grained gabbro, diorite, tonalite and granodiorite of the Lavant Suite (Easton 1988a; Pauk 1989b) and includes the Lavant and Mountain Grove gabbros, as well as several satellite intrusions. The Lavant Gabbro is a composite intrusion consisting of: 1) a slightly older and more voluminous mafic suite, dominated by medium-grained gabbro, which locally shows the presence of igneous layering and several compositionally distinct crosscutting gabbroic and dioritic phases (Figure 19.87); and 2) a slightly younger tonalite-granite phase which consists of tonalite, granodiorite and monzogranite. The younger phase has intruded the gabbro and diorite phases to form intrusion breccias and small intrusive bodies. The tonalites occur mainly in the structurally higher part of the gabbro and the overlying supracrustal rocks, particularly the marbles. The tonalites are associated with the mafic phases of the body and are probably late-stage differentiates of the original magma. Farther south, near Dalhousie Lake, a much larger tonalite-granite body is present, termed the Elphin Pluton (Wu 1984; Wu and Kerrich 1986; Pauk 1989a), and shows similar relationships to the Lavant Gabbro. Lumbers et al. (1990) include the Elphin Pluton within the Skootamatta Suite despite lacking the aeromagnetic or geochemical (Easton, unpublished data) signature typical of the Skootamatta Suite.

The Lavant Gabbro does not produce an extensive contact aureole with the country rocks, although locally, such as at Marble Bluff, it has serpentinized the surrounding marbles. Contact relationships with the country rocks are complex, and inclusions, roof pendants and slivers of all the country rocks are present in the Lavant Gabbro. The Mountain Grove Gabbro (Wolff 1982b, 1985) is similar to the Lavant Gabbro, although it is preserved at somewhat higher metamorphic grade (middle amphibolite *versus* upper greenschist to lower amphibolite). Little chemical data is available from the gabbros of the Sharbot Lake Terrane; what is available is summarized in Easton (1988a).

The Dalhousie Lake amphibolite complex (Pauk 1989a) represents a subvolcanic intrusion and consists of pyroxenite, hornblendite, gabbro, diorite and mafic metavolcanic rocks, generally present as inclusions within the plutonic rocks.

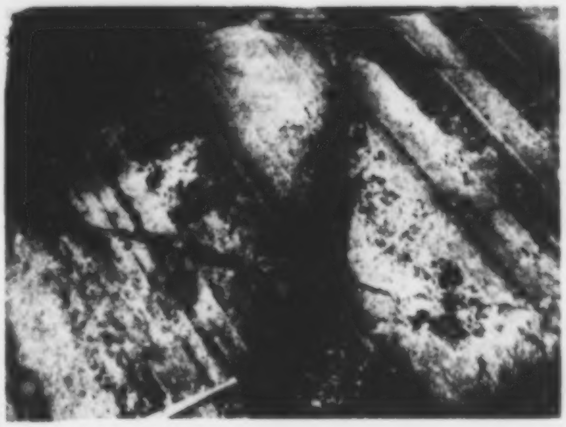


**Figure 19.87.** Igneous layering within the Lavant Gabbro, Sharbot Lake Terrane. Hammer head is 12 cm across.

The Pakenham granodiorite to monzogranite forms a large ovoid pluton located in the northern part of the Sharbot Lake Terrane. The body was intruded as sills with marginal screens of country rocks (e.g., *see* map pattern *of* Reinhardt et al. 1973). The geology and geochemistry of the Pakenham granodiorite to monzonite has not been studied in detail. Lumbers et al. (1990) include it in the Elzevir Suite, however a reconnaissance geochemical survey by the author shows a wide variation in chemical types within the Pakenham pluton and REE patterns atypical of the Elzevir Suite.

In contrast, the Hinchinbrooke Granodiorite Gneiss in the southern part of the terrane is considerably more homogeneous than the Pakenham pluton. Lumbers et al. (1990) also include the Hinchinbrooke Granodiorite Gneiss within the Elzevir Suite, although the main mafic minerals are hornblende and hypersthene (Wolff 1982b), developed through granulite-facies metamorphism (Wolff 1982b). The contact with the surrounding gneisses is faulted, and the pyroxene-bearing mafic and basic migmatitic gneisses (Wolff 1982b) that are the country rock are atypical of the Sharbot Lake Terrane. Wallach (1974) reports a zircon age of 1254 Ma from the Hinchinbrooke Granodiorite Gneiss, suggesting it may be coeval with the Elzevir Suite.

Elzevir Suite plutons are characteristic of the Bancroft, Elzevir and Mazinaw terranes, and the granulite-facies metamorphism is not typical of the remainder of the Sharbot Lake Terrane. It is proposed, herein, that the Sharbot Lake Terrane is a thin thrust sheet that overlies the Mazinaw and Frontenac terranes (*see* Tectonic Summary). If so, the Hinchinbrooke Granodiorite Gneiss and surrounding gneisses may represent a window of the Mazinaw Terrane. This would, in part, explain why Wynne-Edwards (1965, 1967b), in mapping the Tichborne area, included the grey gneisses within the "Frontenac Axis", even though he did note their similarity to gneisses in the Clare River and

Northbrook areas. It would also explain why the Sharbot Lake–Frontenac terrane boundary to the north exhibits a sharp change in metamorphic grade (Easton 1988b) but does not do so near Parham and to the south.

The Barbers Lake Granite is a fluorite-bearing, two mica, alaskitic granite, probably of the Methuen Suite, containing an elevated uranium and thorium content (K.L. Ford, Geological Survey of Canada, Ottawa, unpublished data; Wu 1984; Wu and Kerrich (1986).

## STRUCTURAL HISTORY

The Robertson Lake mylonite zone (RLMZ) was first named by Smith (1958) as the Robertson Lake Shear Zone. Carter (1981) and Jackson (1980) defined the RLMZ as a band of ultramylonites that occurred mainly east of the Addington Granite, although both reported a thin mylonite zone on the west side of the granite. Easton (1988a, 1988f) defined the RLMZ on both the east and west flanks of the Addington Granite in the Lavant–Darling area (Figures 19.88 and 19.89). Easton (1988b) traced the RLMZ north to the Ottawa River and south to Highway 7. The RLMZ dips to the south-southeast at about 40° to 50° along its western margin, and from 15° to 20° along its eastern margin. The character of the mylonites in the RLMZ varies with the adjacent country rocks. Cataclasis in all rocks in the Sharbot Lake Terrane increases toward the RLMZ, which marks the Mazinaw-Sharbot Lake terrane boundary (Easton 1988a), as does carbonate alteration and veining.

The RLMZ extends over 90 km from the Precambrian-Paleozoic boundary north to White Lake and separates rocks that are lithologically, metamorphically and structurally distinct (Easton 1988a, 1988f). The RLMZ is probably a low-angle thrust fault along which rocks to the east have been thrust westward over higher grade rocks. The RLMZ is post-peak metamorphism and is probably similar in age and origin to the Bancroft shear zone (Carlson et al. 1990). It may have been reactivated in the Paleozoic, as the Mt. St. Patrick fault appears to be truncated by the RLMZ, although, the Clyde River fault seems to truncate the RLMZ.

The northern Sharbot Lake Terrane is broken into several crustal blocks by Paleozoic and Tertiary age faults related to the Ottawa–Bonnechere Graben (Easton 1988a). These faults expose various structural levels throughout the Lavant Gabbro and the supracrustal sequence.

## METAMORPHISM

An area of greenschist-facies metamorphism occurs in the northern Sharbot Lake Terrane adjacent to the RLMZ south of White Lake. Metamorphic grade over most of the terrane is lower amphibolite facies, with grade increasing towards the southern and eastern boundaries of the terrane (Easton 1988b; Figure 19.90). Ewert (1977) studied the metamorphism of carbonate rocks located in the northeastern part of the Sharbot Lake Terrane in the Carleton Place map area.

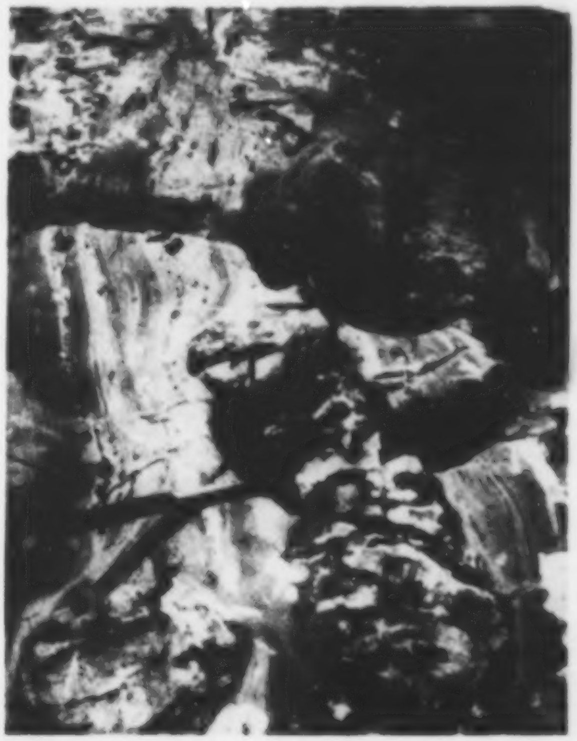


**Figure 19.88.** Mafic mylonite pods within calcitic mylonites of the Robertson Lake mylonite zone. Hammer handle is 30 cm long.

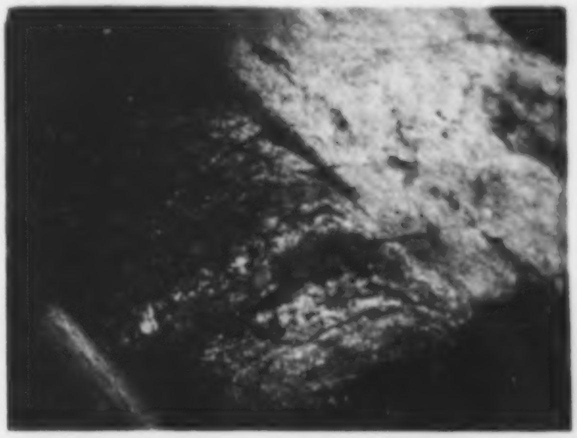


**Figure 19.89.** Mylonitic gabbro within the Robertson Lake mylonite zone. Hammer handle is 30 cm long.

The increase in grade toward the Frontenac Terrane boundary and the presence of straight gneisses in the boundary (Figure 19.91), both suggest that the Frontenac metamorphic event may have affected the terrane boundary and that the terrane boundary formed prior to the end of Frontenac metamorphism, or at roughly *circa* 1170 Ma. Alternatively, the increase in metamorphic grade in the

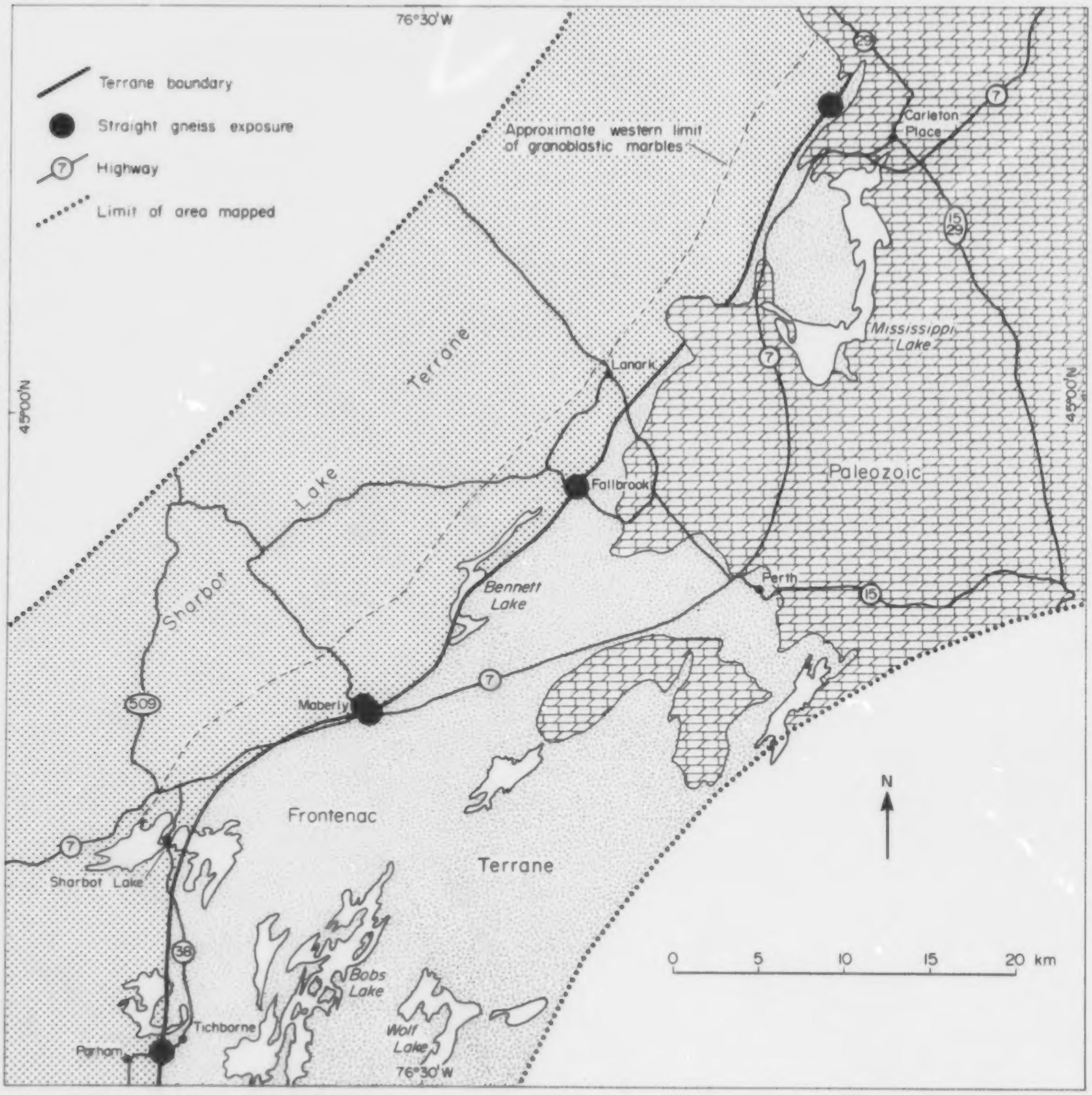
Sharbot Lake Terrane to the southeast and east may simply reflect a regional increase in metamorphic grade within the domain unrelated to Frontenac Terrane metamorphism.

## BASEMENT

No basement rocks have been described from the Sharbot Lake Terrane, as is the case for the Elzevir Terrane.

## GEOLOGIC HISTORY

The Sharbot Lake Terrane was one of active volcanism, of probable Surtseyian type, consisting of the eruption of pillow lava, massive flows, tuffs and the development of cinder and tuff cones of mainly basaltic composition. In part, these eruptions took place in an active carbonate basin. Locally, exhalative activity and wave erosion formed the fine-grained clastic sediments and sulphide-facies ironstones. Farther from the volcanic centre, volcaniclastic debris formed extensive debris aprons around the volcanoes. As volcanic activity waned, carbonate deposition increased and buried the volcanic pile and the debris aprons. Shallow areas saw deposition of dolomites and growth of algal mats, followed by deposition of limestones. Deeper areas accumulated thinly bedded calcite turbidites. The supracrustal sequence was then intruded by gabbros of the Lavant Suite, some phases of which may have been feeders to the volcanic succession. Local hydrothermal alteration and carbonate veining occurred in the volcanic rocks near the gabbro. Tonalites were intruded after the gabbro cooled and differentiated. Subsequently, the



76°30'W
Terrane boundary
Straight gneiss exposure
7
Highway
Limit of area mapped
Approximate western limit of granoblastic marbles
29
7
Carleton Place
15
29
Mississippi Lake
Lanark
7
45°00'N
45°00'N
Terrane
Lake
Sharbot
Paleozoic
Fallbrook
Perth
15
Bennett Lake
7
509
Maberly
Frontenac
Terrane
N
7
Sharbot Lake
38
Bobs Lake
Tichborne
Wolf Lake
Parham
76°30'W
0
5
10
15
20 km

**Figure 19.90.** Frontenac–Sharbot Lake terrane boundary between Tichbourne and Carleton Place (*modified from* Easton 1988b).

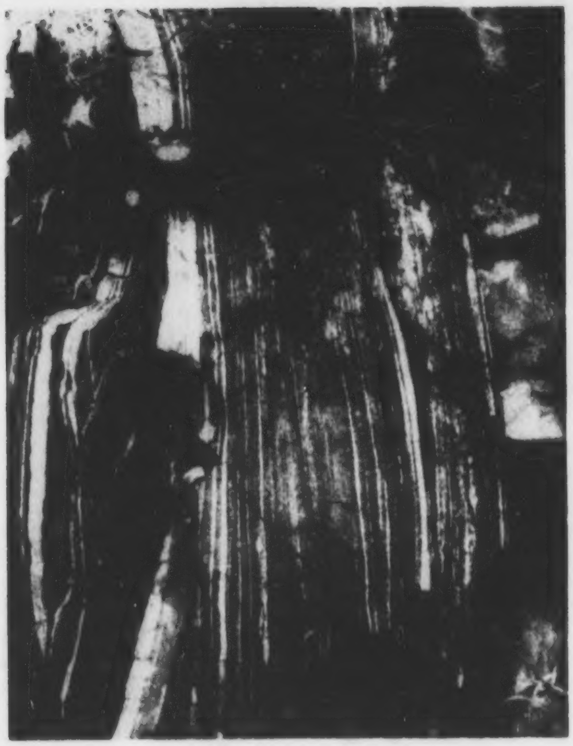


**Figure 19.91.** Straight gneiss marking the Sharbot Lake–Frontenac terrane boundary, Highway 7, near Maberly. Hammer handle is 30 cm long.

compositionally heterogeneous Pakenham intrusion and the fluorine-bearing, Barbers Lake Granite of the Methuen Suite were intruded into the Sharbot Lake Group.

## MINERAL DEPOSITS

### Gold Mineralization

Copper-gold-antimony deposits in the area previously described in detail by Carter et al. (1980), Pauk (1989b) and Carter (1981), and new deposits outlined by Gleeson-Rampton Explorations Limited (Assessment Files Research Office [AFRO]), are generally hosted in ferroan dolomite horizons within the eastern part of the RLMZ (Easton 1988a; LeBaron 1991) derived from the Sharbot Lake Group. Carter (1981) inferred a syngenetic origin, however, subsequent alteration and carbonatization associated with the RLMZ has probably served to concentrate mineralization. Rampton et al. (1986) and Gleeson et al. (1989) have shown that till geochemistry is a useful exploration tool within the RLMZ and the eastern domain.

### Stratiform Zinc Mineralization

Several large zinc occurrences in dolomitic and calcitic marbles are present within the Sharbot Lake Terrane and include: the Long Lake Zinc Mine and related occurrences, the Slave Lake deposit, the Wilkinson occurrence and the Clyde River (Hopetown) occurrence. The Slave Lake and Long Lake deposits are hosted in calcitic marbles preserved as roof pendants in the Hinchinbrooke Granodiorite Gneiss and the Mountain Grove Gabbro (Wolff 1982b; Carter 1984). Wolff (1982b) suggests that the deposits were originally strata-bound, but contact metamorphism remobilized mineralization resulting in deposits with high grades but limited extent. The Long Lake Zinc Mine, discovered in 1896, has been active several times since, most recently from 1973–1974 producing 9467 tons of zinc valued at over $1 million (Wolff 1982b).

The Clyde River occurrence consists of conformable layers and lenses of sphalerite, pyrite and rare galena in dolomitic and siliceous dolomitic marbles (Carter 1984). The deposit was discovered in 1975. The deposit defines a classic ribbon-shaped dispersal train in till (Sinclair 1979; DiLabio et al. 1982, 1987).

### Sulphide Mineralization

Pyritiferous sedimentary rocks associated with metavolcanic rocks and volcaniclastic metasedimentary rocks in the Darling and Sharbot Lake areas, respectively, show no systematic metal enrichment, although local enrichments of gold, copper and zinc do occur (Easton 1988a; Wolff 1985).

### Oxide Mineralization

The Lavant Gabbro has been little studied. It is a differentiated igneous intrusion, is locally layered and shows a considerable compositional range from ultramafic phases to gabbroic anorthosite and tonalite. It may be a worthwhile target for platinum exploration. Several small massive magnetite deposits occur in the Lavant Gabbro and include the Yuill iron mine, the Darling and Lavant deposits (Carter et al. 1980) and several aeromagnetic anomalies reported by Gleeson-Rampton Explorations Limited (AFRO)(Easton 1988a). These deposits all occur near the margin of the Lavant Gabbro and are mainly magmatic types.

Iron skarn deposits are likely present in marbles in the contact aureole of the Lavant and Mountain Grove gabbros, as well as smaller satellite intrusions in the area. Industrial minerals such as wollastonite and talc can also be expected in the same setting.

### Industrial Minerals

Two major stone quarries have operated in the past in the northeastern Sharbot Lake Terrane. The Tatlock Omega (Blue) Marble Quarry operated between 1962 and 1971 producing blue, white, pink and buff-coloured marble for decorative stone and terrazzo uses (Storey and Vos 1981a). Between 1963 and 1971, Anglestone Limited operated a marble quarry 750 m southwest of the blue marble quarry (Storey and Vos 1981a). Since 1977, this area has been the site of a large quarry operation which supplies white crushed marble (95 to 97 brightness) to the Steep Rock Resources Inc. calcite plant in Perth, Ontario for use in fine and ultra-fine calcite products (Easton et al. 1986). A third quarry operated nearby on Highway 511 at Marble Bluff in a serpentized marble band but produced very little stone

(Peach 1958). The Sharbot Lake Terrane has considerable potential for further marble production.

Other industrial commodities reported from the area include marl along the west shore of Joe Lake, and possibly in the area of Clay and Little Green Lakes. Quartz-tremolite horizons are present within many of the dolomite marbles, although no massive tremolite deposits have been located, other than the Ram Petroleums Limited tremolite quarry (Pauk 1989a) located in a roof pendant in the Addington Granite within the RLMZ. No graphite prospects have been reported from the terrane.

# FRONTENAC TERRANE

## Introduction

Almost all of the Frontenac Terrane has been mapped at 1:50 000 scale or better (Baker 1916, 1923; Wright 1923; Wilson 1930; Harding 1951; Dugas 1950; Wilson and Dugas 1961; Wynne-Edwards 1959, 1962, 1963a, 1965, 1967a). Hewitt (1964a) and Kingston et al. (1985) have produced 1:126 720 and 1:125 000 scale compilation maps, respectively, covering most of the Frontenac Terrane. Although contacts and descriptions of major rock units are generally correct, mapping in the terrane has not been done in the post-plate-tectonic era, nor in light of major advances in structural geology and stratigraphy in highly deformed regions. As a result, our understanding of the geology of the Frontenac Terrane lags behind that of the rest of the Central Metasedimentary Belt.

Rocks from the Frontenac Axis have long been discussed separately from those of the rest of the Central Metasedimentary Belt (e.g., Baker 1916, 1923; Wright 1923, Wilson 1930; Harding 1951; Wynne-Edwards 1959, 1962, 1963a, 1965, 1967a), although they were generally considered to be part of the Grenville Supergroup. Wynne-Edwards (1967a) defined the Frontenac Axis as those Precambrian rocks lying between the Clare River synform and the St. Lawrence River. Moore's (1982) boundary lay farther east than that of Wynne-Edwards, generally coincident with the boundary shown in Figure 19.90. Corriveau (1989, 1990) and Corriveau et al. (1990) suggested that the Frontenac Terrane, proper, lies south of the Rideau–Canoe Lake fault, based on the distribution of late tectonic to posttectonic syenitic intrusions in the Central Metasedimentary Belt. The terrane is here defined based on the boundaries of Moore (1982) placing the boundary of the Frontenac Terrane as shown on Figure 19.90.

## Basement(?) to the Supracrustal Sequence

Unlike most of the rest of the Central Metasedimentary Belt, there have been several suggestions that older basement underlies the supracrustal sequence in the Frontenac Terrane (e.g., Wynne-Edwards 1967a, 1967b; Moore 1982). The older suggestions were based on the platformal sedimentary sequence in the region, as well as on structural and compositional grounds.

Recent isotopic studies support the concept of older crustal material of probable Paleoproterozoic age underlying the Frontenac Terrane. McLelland, Chiarenzelli et al. (1988) report a zircon age of 1415±6 Ma from a leucocratic orthogneiss on Wellesley Island in the Adirondack Lowlands. The orthogneiss intrudes previously deformed metasedimentary rocks, yet is cut by younger, Rockport-type leucogranites. Quartzite xenoliths from the Wellesley Island orthogneiss contain zircons of 1807 and 1992 million year age (McLelland, Chiarenzelli et al. 1988). In addition, Nd-Sm model ages from paragneisses throughout the Frontenac Terrane (Marcantonio et al. 1990) are all in the range 2045 to 1560 Ma, indicating derivation from a crustal source of Paleoproterozoic age. Marcantonio et al. (1990) and Shieh (1985) both note that crustal melting and contamination played a role in generation of the plutons found throughout the Frontenac Terrane. In addition, Kinsman and Parrish (1990) found that detrital zircon ages from the quartzites in the Frontenac Terrane ranged from 1.3 to 3.2 Ga, with clusters of ages at 1301±13, and between 1745 and 1892 Ma, consistent with the ages reported by McLelland, Chiarenzelli et al. (1988). This indicates the existence of a source region capable of supplying an abundance of *circa* 1300 and *circa* 1800 million-year-old detritus. Such a source area may be local (i.e., within the Frontenac Terrane), or more distal (probably to the south, as quartzite units become thicker and more abundant to the south). To date, crust older than 1415 Ma has not been positively identified in the Frontenac Terrane.

## Frontenac Supergroup

### INTRODUCTION

The supracrustal rocks of the Frontenac Terrane (Axis) are herein distinguished as the Frontenac Supergroup (*see* Table 19.7). These rocks differ from the Grenville Supergroup in their lack of volcanic and volcaniclastic rocks and the long-noted abundance of quartzite units. In addition, as discussed below, sedimentation within the Frontenac Terrane may have occurred at times very different from sedimentation elsewhere in the Central Metasedimentary Belt. The stratigraphy of the area has not been studied in detail since the work of Wynne-Edwards (Wynne-Edwards 1967a), and thus no subdivisions are proposed herein for the Frontenac Supergroup, although undoubtedly several could be established based on the stratigraphy outlined by Wynne-Edwards (1967a) (*see* Figure 19.55a).

The stratigraphic framework used here follows that of Wynne-Edwards (1967a), as illustrated in Figure 19.55a. Each of the major subdivisions would correspond to a "group" but are not defined as such given the current uncertainty about stratigraphic relations and relative and absolute ages of supracrustal rocks with the Frontenac Terrane. Wynne-Edwards (1967a) stratigraphy is based on several important assumptions, namely: 1) continuously traceable lithologic contacts represent single stratigraphic horizons; 2) marble units represent reliable datum horizons; 3) a limited number of top determinations from quartzite horizons are sufficient to determine regional stratigraphic-facing directions; 4) that all gneiss-marble contacts are normal sedimentary contacts (i.e., not faulted); and 5) there are 2 marble horizons which are in contact at Morton. If the latter assumption is incorrect, as noted by Wynne-Edwards (1967a), it would not only complicate the structural interpretation of the area, but would also imply that there is only a single marble unit that lies above a single gneiss unit. It is noteworthy that detrital zircon studies of Kinsman (1990) appear to support the stratigraphy proposed by Wynne-Edwards (1967a) as outlined in Figure 19.55a. Wynne-Edwards (1967a) estimated the sedimentary succession to be about 5000 m thick, with neither the top nor the base recognized. Tectonic interleaving is probably present in the area.

## LOWER GNEISS AND GRANULITE

Within the Frontenac Terrane, there are a variety of paragneisses, distinct from the marble-quartzite sequence (*see* stratigraphic sequence, Figure 19.55a). Most are migmatitic. Wynne-Edwards (1967a) notes that the proportion of granitic leucosome in the gneisses varies considerably, and he used distinctive minerals to distinguish different varieties of paragneiss. Varieties include *lit-par-lit* gneisses, biotite gneisses, garnet gneisses, cordierite gneisses and pyroxene gneisses and are fully described in Wynne-Edwards (1967a). Most are quartzofeldspathic in composition, and the gneisses were probably derived, for the most part, from wackes, siltstones and marls.

In addition to the gneisses, Wynne-Edwards (1967a) describes massive, quartzofeldspathic granulites distinguished from the gneisses by lower mafic mineral contents and their weak to absent foliation. Like the gneisses, the granulites are commonly migmatitic. The granulites contain up to 40% quartz, as well as oligoclase ($\mathrm{An}_{[illegible]-35}$) and antiperthite. Wynne-Edwards (1967a) noted their similarity to feldspathic quartzites higher in the stratigraphic sequence.

## MAJOR MARBLE

Defined by Wynne-Edwards (1967a), the Major Marble consists of coarse (average grain size of 3 mm) calcitic marble, com-monly carrying minor graphite, phlogopite, diopside, scapolite, apatite, tourmaline and pyrite. Layers and blocks of calc-silicate gneiss, generally rusty, are common in the unit. The type areas are near Devil and Opinicon Lake in the Westport map area (Wynne-Edwards 1967a) and near Sand Lake.

Although Wynne-Edwards (1967a) reported no reliable field criteria for distinguishing between the Major Marble and the stratigraphically higher Marble-Quartzite unit, all major graphite occurrences (*see* Graphite section below) in the Frontenac Terrane occur in the Major Marble, west of, and within 10 km of the Rideau–Canoe Lake fault. As other major faults occur throughout the Frontenac Terrane without associated graphite mineralization, this may indicate that there is a stratigraphic, in addition to a structural, control on graphite distribution in the Frontenac Terrane.

## MAIN GNEISS UNIT

Defined by Wynne-Edwards (1967a), the Main Gneiss Unit consists of quartzofeldspathic diopside granulite, layered pyroxene gneiss and amphibolite and quartz-biotite-feldspar gneiss with thin quartzite and garnet gneiss layers in the west. Farther east, quartzite and feldspathic quartzite become dominant, along with biotite-cordierite-sillimanite gneiss and dark gneiss with garnet-quartz porphyroblasts. The type areas are north of Devil Lake, near Hamilton Lake, and the Morton syncline. No field criteria were noted by Wynne-Edwards (1967a) for distinguishing the Lower Gneisses from the Main Gneiss Unit and, as in the case of the marble units, the stratigraphic division is made primarily on structural interpretation *versus* lithologic characteristics.

## UPPER MARBLE AND QUARTZITE

This unit is present in the eastern part of the Frontenac Terrane and consists of calcitic and minor dolomitic marbles, quartzite, garnet-sillimanite gneiss and a "quartzite-limestone" transition rock (Wynne-Edwards 1967a), which consists of centimetre-scale interlayered bands of quartzite and marble. This unit is only present in the southeastern part of the terrane, with type areas near Lyndhurst, Charleston Lake, in the Red Horse dome, and in the Mountain anticline. Wynne-Edwards (1967a) distinguished the Upper and Lower marbles on structural criteria and reported no reliable criteria for distinguishing the 2 types in the field. Stromatolites have not been reported from the Frontenac Terrane marbles, although the author observed stromatolites in the possibly correlative upper marble sequence in the Adirondack Lowlands at Balmat, New York.

## MIGMATITES

Migmatites occur throughout the sedimentary succession in the terrane, particularly within the quartzofeldspathic gneisses. Leucosomes are commonly granitic, or quartz-potassium feldspar aggregates. Wynne-Edwards (1967a) also distinguished injection migmatites near plutons of the monzonite-granite-syenite suite. These injection migmatites occur in zones up to several kilometres wide around plutons. Red and white pegmatites are commonly found in association with the migmatites, with white pegmatites being found only in areas where marble is present.

## ARE TWO SEDIMENTARY SEQUENCES PRESENT WITHIN THE FRONTENAC TERRANE?

Geochronologic studies on Frontenac Terrane supracrustal rocks both in Ontario and in the Adirondack Lowlands in New York present conflicting results on the age of Frontenac Supergroup sedimentation. McLelland, Chiarenzelli et al.'s (1988) date on the Wellesley Island leucogranitic gneiss, which includes metasedimentary xenoliths, implies sedimentation occurred prior to *circa* 1420 Ma. The youngest detrital zircons reported by Parrish (1990) and Kinsman and Parrish (1990) from a quartzite from the lower part of Wynne-Edwards' (1967a) stratigraphic succession were 1415±2 million years in age, consistent with the data from Wellesley Island. The youngest detrital zircons reported by Parrish and Kinsman and Parrish (1990) from a quartzite from the upper part of Wynne-Edwards' (1967a) stratigraphic succession near Charleston Lake, are 1301±13 million years in age, implying sedimentation occurred subsequent to 1300 Ma.

One explanation for this variance in ages is that rock suites of 2 ages may be present in the Frontenac Supergroup. Nd-Sm model ages reported from Marcantonio et al. (1990) are all from paragneisses within the Frontenac Supergroup, not marbles or quartzites. This data places a maximum age on deposition of these paragneisses at *circa* 1750 Ma, with a minimum age of *circa* 1420 Ma based on the age of the Wellseley Island leucogranite gneiss. Deposition of some quartzite and marble units took place after *circa* 1300 Ma, based on the detrital zircon study of Kinsman and Parrish (1990). Paleolatitude data of Irving (1979) suggests the area lay in a favourable climate for carbonate deposition between 1450 and 1250 Ma and 1800 and 1750 Ma, assuming that the Frontenac Terrane was not significantly distal from Laurentia during this time span.

If 2 sedimentary packages are present within the Frontenac Terrane, they cannot be outlined based on existing mapping, although the bulk of the quartzite and marble units probably belong to the younger period of sedimentation. In summary, it is clear that the stratigraphy within the Frontenac Supergroup, as outlined by Wynne-Edwards (1967a), needs revision.

### FACIES(?) VARIATION ACROSS THE FRONTENAC TERRANE AND DEPOSITIONAL ENVIRONMENTS(?)

Wynne-Edwards (1967a) reports that quartzite becomes more abundant from northwest to southeast across the Westport area, and that quartzite becomes the dominant metasedimentary unit along the St. Lawrence River.

The sedimentary succession in the Frontenac Terrane suggests a platformal environment, probably in relatively shallow water given the abundance of carbonates and thick quartzite units in the southeast. This facies variation may indicate a deepening of the shelf to the west and northwest, with a very shallow-water or shoreline environment lying to the southeast. The presence of evaporite units in the Adirondack Lowlands near Balmat, New York (deLorraine and Dill 1982) is consistent with such an environment, as is the apparent abundance of stromatolites in the lowlands. The distribution of dolomitic rocks in the Frontenac Terrane and Adirondack Lowlands, based on limited data, is also consistent with this model. The majority of marbles within the Frontenac Terrane are calcitic, with dolomitic marbles becoming slightly more prevalent to the southeast. Dolomitic marbles are common in the lowlands, particularly in the Upper Marble (deLorraine and Carl 1986). Stromatolites are commonly associated with dolomitic marbles, an association that may be due to dolomitization in the near-shore, shallow-water environments conductive to stromatolite growth, and diagenesis and dolomitization prior to lithification. Thus, the paucity of stromatolites and dolomites from the Frontenac Terrane in Ontario may be stratigraphically controlled.

### CORRELATION WITH THE GRENVILLE SUPERGROUP

Correlation between the Grenville Supergroup and the Haliburton Group in the Elzevir and Bancroft terranes, respectively, and the Frontenac Supergroup cannot be established. As outlined above, supracrustal rocks within the Frontenac Terrane may have been deposited in 2 intervals. If so, then only the younger sedimentary episode may be roughly time-correlative with the other 2 supergroups, and even then, deposition of the Frontenac Supergroup may have occurred over a much longer time span (*see* Tectonic Summary section below).

## Plutonism

### DIORITE-GABBRO AND ANORTHOSITE SUITE (UNKNOWN AGE)

Small bodies of diorite, gabbro, gabbroic anorthosite and anorthosite occur throughout the Frontenac Terrane, although the gabbroic anorthosite and anorthosite bodies occur mainly in the western part of the Frontenac Terrane near the boundary with the Sharbot Lake Terrane. Gabbro bodies include the Newboro Gabbro, the Wolfe Lake metagabbro, the Westport Gabbro and the Troy Lake Gabbro (Wynne-Edwards 1967a). Gabbroic anorthosite and anorthosite bodies, possibly correlative with the anorthosite suite of the Elzevir Terrane, include the Lanark–Oso meta-anorthosite and the Tichbourne, Victoria, Blessington, St. Andrew and Island Lake gabbroic anorthosites, with the first 3 constituting the largest bodies. Harrison (1944) studied most of the gabbroic anorthosite bodies in detail.

Mineralogy of the gabbros consist dominantly of augite and plagioclase ($An_{30-50}$). The latter commonly having a bluish to a lilac tinge in the cores of the gabbroic anorthosite bodies. Hypersthene, biotite, apatite and green spinel are common accessories. The average gabbro contains 60% andesine, 25% pyroxene, 5% biotite and 10% magnetite and ilmenite (Wynne-Edwards 1967a). Magnetite and ilmenite are abundant constituents of the Newboro, Tichborne and Troy Lake gabbros.

None of these intrusions have been dated. Wynne-Edwards (1967a) notes that the gabbros have mineral assemblages compatible with the surrounding gneisses, indicating that both were metamorphosed under the same conditions. This places a minimum age of *circa* 1170 Ma on the bodies (Mezger et al. 1990). This also makes the gabbros older, even if only slightly older, than the granitoid rocks in the area. Carter (1984), however, suggested that the Newboro Gabbro was unmetamorphosed. Wynne-Edwards (1967a) refers to the Newboro Gabbro as "unaltered" but considers it to be metamorphosed. If Carter (1984) is correct, then the Newboro may be related to the Skootamatta Suite and would be younger than 1170 Ma. Frontenac Terrane gabbros are dominated by the presence of ilmenite and titaniferous magnetite, whereas magnetite is predominant in Elzevir Terrane gabbros and gabbroic anorthosites.

Whether the Frontenac Terrane gabbros are correlative with the Lavant, Killer Creek, or anorthosite suites of the Elzevir Terrane is not presently known.

## MONZONITE-SYENITE-GRANITE SUITE (1180 TO 1165 Ma)

As noted in the overview on Central Metasedimentary Belt magmatism, the plutonic suite of the Frontenac Terrane is distinct. Rocks from this suite have been studied by numerous workers, including Wynne-Edwards (1967a), Currie and Ermanovics (1971), Shieh (1985), Marcantonio et al. (1990) and Lumbers et al. (1990).

The suite consists mainly of massive syenite and granitic rocks largely unaffected by deformation and recrystallization. The plutons cut mylonitic fabric in their host rocks, but they locally contain post-Grenvillian mylonitic zones where the rocks have been crushed and recrystallized during brittle deformation (e.g., Perth Road mylonite zone, Barclay 1985, Currie and Ermanovics 1971). Rocks of the suite are rich in mesoperthite, sodic oligoclase and biotite, and many plutons contain minor clinopyroxene. Monzonite and syenite plutons are equally abundant. Syenogranite and monzogranite are minor phases in most plutons, with the exception of the monzogranitic Lyndhurst and Battersea bodies. Gabbro and diorite xenoliths are common in felsic plutons of this suite.

Wynne-Edwards (1967a) and Sauerbrei (1966) divided rocks of the suite into 2 types: 1) the Frontenac type, occurring as discrete plutons of syenite to granite; and 2) the Rockport type, mainly biotite granite and monzonite intimately associated with migmatites. Plutons of the Frontenac type include the Lyndhurst Granite, the Gananoque Syenite, the Perth Road Monzonite, the Crow Lake Monzonite, the Battersea Granite, the South Lake Monzonite, the Charleston Lake Syenite and the Brockville Monzogranite. The Rockport Granite is the only major Rockport-type pluton. Plutons of both types alternately intrude and include one another, although Wynne-Edwards (1967b) notes that the Rockport Granite is generally conformable with the country rocks. Many plutons have now been dated, and the Rockport-type pluton appears to be slightly older, having an age of $1173^{+4}_{-3}$ Ma (Rockport Granite, van Breemen and Davidson 1988b). Frontenac-type plutons are slightly younger, with ages of 1162±3 Ma (Gananoque Syenite, van Breemen and Davidson 1988b), 1166±3 Ma (Lyndhurst Granite, Marcantonio et al. 1990), 1166±3 Ma (Perth Road Monzonite, Marcantonio et al. (1990), 1165±4 Ma (Battersea Granite, Marcantonio et al. 1990) and 1176±2 Ma (Crow Lake Monzonite, Marcantonio et al. 1990). Nd model ages of the plutons studied by Marcantonio et al. (1990) (Battersea, Lyndhurst, Perth Road, Crow Lake, Gananoque and South Lake plutons) range from 1.34 to 1.48 Ga, notably older than their intrusion ages, suggesting crustal contamination of the magmas.

Chemical analyses of the plutons are presented by Marcantonio et al. (1990) and Lumbers et al. (1990), who classify them as I-type plutons. All are diopside normative except for the Crow Lake Monzonite. Marcantonio et al. (1990), using the Pearce et al. (1984) classification scheme for granitoids, found that the monzonite-syenite-granite suite fell into the within-plate granite field.

Shieh (1985) conducted an oxygen isotope study on 9 plutons of this suite and found $^{18}O$ values of *circa* +14‰ (Battersea, Lyndhurst, Perth Road, Crow Lake, Gananoque and South Lake plutons). Marcantonio et al. (1990) reported similar $^{18}O$ values for the Battersea, Lyndhurst, Perth Road, Crow Lake, Gananoque and South Lake plutons. Marcantonio et al. (1990) also reported $\varepsilon_{Nd}{}^{t}$ values of +1.4 to +3.1 for these plutons. Shieh (1985) suggested that these plutons, all located south of the Rideau–Canoe Lake fault, were derived by the melting of high $^{18}O$ Frontenac Supergroup paragneisses. However, a Rb-Sr age for the Lyndhurst Granite of 1073±49 Ma (Krogh and Hurley 1968) has a low $^{87}Sr/^{86}Sr$ initial ratio of 0.704, suggesting a primary mantle source. This suggestion is at variance with the conclusion of Shieh (1985) but is consistent with the I-type chemistry of the suite. Marcantonio et al. (1990) favoured a model of granitoid petrogenesis involving mixing of a basic crustal source, with contamination by surrounding metasedimentary rocks having a Nd model age of 1790 Ma, and interaction with marble-derived fluids.

## SKOOTAMATTA SUITE (1090 TO 1075 Ma)

As noted in the overview on Central Metasedimentary Belt magmatism, this plutonic suite is present in all terranes within the Central Metasedimentary Belt, however, it has not been reported in the Frontenac Terrane south of the Rideau–Canoe Lake fault. Rocks from this suite have been studied by numerous workers, including Wynne-Edwards (1967a, 1967b), Corriveau (1989, 1990), Corriveau et al. (1990), Shieh (1985), Marcantonio et al. (1990) and Lumbers et al. (1990).

In the Frontenac Terrane, plutons of this suite occur in a cluster on the north shore of Rideau Lake and include the Westport, Wolfe lake and Rideau Lake bodies. All are monzonites containing little or no quartz. Plagioclase is generally $An_{10-15}$, and the potassium feldspar is mostly microcline with some orthoclase. Biotite is the dominant mafic mineral, although hornblende is present in the Westport Monzonite. Titanite, zircon, apatite, calcite and magnetite are ubiquitous accessory minerals.

Chemical analyses of the plutons are presented by Marcantonio et al. (1990) and Lumbers et al. (1990) who classify them as I-type plutons. All are diopside normative. Lumbers et al. (1990) and Marcantonio et al. (1990) both note a similarity in chemistry between rocks of these suite and the monzonite-granite-syenite suite, however, the syenite suite plutons have less $SiO_2$ and CaO, but more $Al_2O_3$, $K_2O$ and $Na_2O$ than the monzonite-syenite-granite suite. Marcantonio et al. (1990), using the Pearce et al. (1984) classification scheme for granitoids, found that the syenite suite plutons in the Frontenac Terrane fell into the within-plate granite field, as did the monzonite-syenite-granite suite plutons.

Shieh (1985) conducted an oxygen isotope study on the 3 plutons of this suite in the Frontenac Terrane and found $^{18}O$ values of *circa* +10‰. Marcantonio et al. (1990) reported similar $^{18}O$ values, as well as $\varepsilon_{Nd}{}^{i}$ values of +2.7 to +3.4 for these plutons. Shieh (1985) suggested that the plutons were derived by the mixing of melts of Frontenac Supergroup paragneisses and basement gneisses (i.e., 2 distinct crustal melt sources). However, a Rb-Sr age for the Westport Monzonite of 995±38 Ma (Krogh and Hurley 1968) has a low $^{87}Sr/^{86}Sr$ initial ratio of 0.7054, suggesting a primary mantle source. This suggestion is at variance with the conclusion of Shieh (1985) but is consistent with the I-type chemistry of the suite. Marcantonio et al. (1990) favoured a model of granitoid petrogenesis involving mixing of a basic crustal source, with contamination by surrounding metased-imentary rocks having a Nd model age of 1790 Ma and interaction with marble-derived fluids, similar to the origin of the monzonite-syenite-granite suite plutons.

Wynne-Edwards (1967a) reports biotite dike rocks associated with the Westport Monzonite. These dikes are probably related to the intrusion and have been reported from other bodies of this suite in the Elzevir Terrane (Easton, unpublished data).

## Metamorphism

### TIMING OF METAMORPHISM

The Frontenac Terrane apparently has a monocyclic metamorphic history in contrast to the polycyclic histories present throughout the rest of the Grenville Province (*see* Figure 19.14). Mezger et al. (1990) dated this metamorphic event at *circa* 1170 Ma using U-Pb zircon, titanite and monazite ages. This metamorphic event is coincident with intrusion of the monzonite-syenite-granite suite plutons. Postmetamorphic cooling was rapid, as Ar-Ar ages on hornblendes from the Frontenac Terrane are all between 1125 and 1105 Ma (Cosca 1989), indicating cooling from roughly 800° to 500°C in less than 50 million years. Lonker (1980) noted the abundance of disequilibrium textures in the granulites, perhaps indicating widespread, but localized retrogression.

### GRANULITE-FACIES METAMORPHISM

As noted in the introductory section on metamorphism, granulites of the Frontenac Terrane were metamorphosed under lower pressure conditions than the Central Gneiss Belt granulites. Frontenac Terrane granulites are similar to the metasedimentary-type granulite terrane of Percival (1988) in the Superior Province, in having almost congruent plutonism (1175 to 1165 Ma, Marcantonio et al. 1990, van Breemen and Davidson 1988b) and metamorphism (4 to 6 kb, 700° to 800°C at *circa* 1170 Ma, Mezger et al. 1990), sedimentation to within possibly 20 million years of plutonism and rapid uplift and cooling (Flinton Group contains detritus of Frontenac plutons at 1157±3 Ma, Kinsman and Parrish 1990, and Cosca (1989) reports Ar-Ar ages on hornblendes of 1120 to 1110 Ma indicating cooling below roughly 500°C by that time). Wynne-Edwards (1967a, 1967b) and Carmichael et al. (1978, 1987) show metamorphic conditions peaking around Westport, Ontario (*see* Figure 19.14), and decreasing to the north and south. Lonker (1980), however, suggests that the highest metamorphic grade is found north of Gananoque, and that no culmination exists near Westport. The granulites of the Frontenac Terrane have been the subject of considerable research (*see* Table 19.1).

## Structure

The Frontenac Terrane has had a complex structural history as summarized by Wynne-Edwards (1963b, 1967a, 1967b). Wynne-Edwards (1967a) also recognized a number of structural domains within the Westport area. At least 2 major periods of folding occurred in the area. Early folds have steeply dipping axial planes trending northeasterly. Later folds have axes lying on steeply dipping planes trending east-northeast. These 2 fold patterns result in a complex interference pattern throughout the Frontenac Terrane. Wynne-Edwards (1967a) regarded the 2 phases of folding as part of a protracted deformation event, rather than separate deformational events. Carmichael et al. (1987) noted that although the kinematic interpretation of folding made by Wynne-Edwards (1963b, 1967a) was debatable, the overall fold geometry was probably correct. However, they did suggest that early thrusts might be present in the sequence, with obvious implications for stratigraphic analysis in the area. Easton and Ford (1991) note that the structural history of the Mazinaw and Frontenac terranes and the Adirondack Lowlands are similar, and all of these areas contain documented early thrusts.

Extensional mylonite zones occur throughout the Frontenac Terrane. The most prominent of these are shown on Figure 19.19 and include the Rideau–Canoe Lake fault and the Perth Road mylonite zone (Barclay 1985; Carmichael et al. 1987). The Perth Road mylonite zone does not affect Paleozoic rocks (Carmichael et al. 1987) and contains lower amphibolite mineral assemblages, indicating that it postdates peak metamorphism. Shear-sense indicators give no consistent sense throughout the zone (Barclay 1985; Carmichael et al. 1987).

## Adirondack Lowlands-Frontenac Terrane Relations

### INTRODUCTION

A metamorphic grade change occurs at the St. Lawrence River, separating granulite-facies rocks of the Frontenac Terrane in Ontario from upper amphibolite-facies rocks of the Adirondack Lowlands in New York State. This metamorphic change may be related to downdropping of the Adirondack Lowlands along the St. Lawrence Graben System, thereby preserving a slightly lower grade, structural section in the lowlands.

### STRATIGRAPHY

As shown in Figure 19.55b, the stratigraphic successions in the 2 regions are similar. However, more than one interpretation of the "stratigraphic" relations in the Adirondack Lowlands is possible. Because of these similarities, work in the Adirondack Lowlands can be used to better constrain the geologic history of the Frontenac Terrane, which is defined to include the Adirondack Lowlands. deLorraine and Carl (1986) review these various stratigraphic interpretations. The lowermost Hyde School Gneiss in Figure 19.55b has been variously interpreted as a felsic volcanic succession or a granitoid intrusion. The Popple Hill Gneiss has also been considered to be a dacitic volcanic or a granodiorite intrusion. Many lithologic contacts have been assumed to be conformable, as in the Frontenac Terrane, although this assumption has also been questioned.

Some rock types in the lowlands have not been reported from the Frontenac Terrane. These include tourmalinites, anhydrite-bearing rocks and stromatolitic dolomites. Most of these rock types are found only in the Upper Marble, and the tourmalinites are interpreted to have formed from boron-rich brines present in an evaporitic sequence (Brown and Ayuso 1985). As noted elsewhere in this section, the lack of a shallow-water, evaporitic environment found in the Balmat–Edwards area in the Adirondack Lowlands is not reported from Ontario and may reflect regional facies variations within the Frontenac Terrane.

### PLUTONISM

Both Frontenac- and Rockport-type plutons are reported from the Adirondack Lowlands. Correlation is difficult since the terminology used by Wynne-Edwards (1967a, 1967b, 1967c) has not been used commonly in the lowlands. Rock types absent from the Frontenac Terrane in Ontario include alaskitic gneisses such as the Hyde School Gneiss, tonalite-granodiorites of the Popple Hill Gneiss and granites of the Antwerp and Rossie type. McLelland et al. (1992) report a U-Pb zircon age of 1230±33 million years for the Hyde School Gneiss, suggesting correlation with the Methuen Suite. In contrast, the Antwerp–Rossie granitoids have zircon ages of 1183±7 million years (McLelland et al. 1992), similar to Rb-Sr ages of 1197±53 and 1160±42 Ma (Carl et al. 1990) suggesting correlation with the Gananoque Suite.

### METAMORPHISM

The Adirondack Lowlands was metamorphosed to upper amphibolite facies. Pressure-temperature conditions summarized in deLorraine and Carl (1986) indicate conditions of 600° to 650°C and 6 to 6.5 kb. Temperatures in the Frontenac Terrane granulites are roughly 150° to 200°C higher.

### STRUCTURE

deLorraine and Carl (1986) note 4 phases of folding in the Adirondack Lowlands. The first phase resulted in northeasterly trending isoclinal folds, and migmatitization may have occurred at this time. Second-phase, fold axial surfaces trend northeasterly, as well, but vary in trend and plunge. Thrusting accompanied the second phase folds (e.g., deLorraine and Dill 1982), and locally overturned folds were developed. Third phase folds trend north-northeast, are tight to isoclinal and are upright or overturned to the southeast. Fourth phase folds are gentle, open folds that trend northwesterly, and interference between the second and fourth phase folds results in the dome and basin pattern found in the region. This is a more complex pattern than outlined by Wynne-Edwards (1967a), but the phase 1 and 2 folds of the lowlands may be transposed in Ontario as a result of the higher degree of metamorphism.

### GEOCHRONOLOGY

McLelland, Chiarenzelli et al. (1988) argue that all metasedimentary units in the lowlands may all be older than 1420 Ma, the age of the Wellesley Island leucogranite gneiss. McLelland, Chiarenzelli et al. (1988) also report a U-Pb zircon age of 1284±7 Ma for alaskitic gneisses of the Gouverneur dome (*see* Figure 19.55b, Hyde School Gneiss), leading them to conclude that the Hyde School Gneiss intrudes metasedimentary rocks in the area, rather than being the basal unit in the stratigraphic sequence. This of course assumes all metasedimentary rocks in the area are older than *circa* 1420 Ma. McLelland et al. (1991, 1992) provide additional evidence that the Hyde School Gneiss contact metamorphosed the surrounding metasedimentary rocks, further suggesting that metasedimentary rocks in the lower marble are older than *circa* 1300 Ma. The age of metamorphism in the lowlands is constrained by a Rb-Sr age of 1154±19 Ma (McLelland, Lochhead and Vyhnal 1988; Hudson et al. 1986) from a garnet-sillimanite tectonite unit at the base of the Popple Hill Gneiss (also known as Major Gneiss) and a U-Pb monazite age of 1137±1 Ma from a Frontenac-type granite (McLelland, Chiarenzelli et al. 1988). These ages are consistent with ages from the Frontenac Terrane (Mezger et al. 1990). The Popple Hill Gneiss has a Rb-Sr age of *circa* 1300 Ma (Hudson et al. 1986; McLelland, Lochhead and Vyhnal 1988) and thus, is part of the early tonalite magmatic suite found from the Central Gneiss Belt, the northern Central Metasedimentary Belt, through to the Adirondack Mountains and Grenvillan inliers in Vermont. Rocks of this age have not yet been reported from the Frontenac Terrane in Ontario.

## Mineral Deposits

Many mineral deposit types related to the Central Metasedimentary Belt in regional metallogenic studies (e.g., Sangster and Bourne 1982; Carter 1984) are absent from the Frontenac Terrane. In part, this reflects the absence of volcanic rocks from the Frontenac Terrane, but it may also be a consequence of the fact that the sedimentary and plutonic succession within the Frontenac Terrane is very different from the Bancroft and Elzevir terranes. Of those deposits that are present, many are concentrated in the northern and western parts of the terrane north and west of the Rideau–Canoe Lake fault, in a possibly deeper water part of the Frontenac depositional platform.

### STRATIFORM ZINC DEPOSITS

Only 1 small zinc occurrence is known from the Frontenac Terrane, the 30 Island Lake deposit present about 10 km east of the Frontenac–Elzevir terrane boundary (Harding 1951; Malczak et al. 1985). It is in dolomitic marbles adjacent to a gabbroic intrusion, and, in some ways, resembles the Sharbot Lake Terrane zinc deposits found in marble roof pendants in gabbroic intrusions (Carter 1984). Thus, the deposit is probably better regarded as a mineralized skarn, rather than a true stratiform deposit.

The best known, largest and the longest commercially producing zinc deposits from the Grenville Province are the Balmat–Edwards district zinc deposits, located in the Adirondack Lowlands. Over 23 million tons of ore grading 9% Zn and 0.5% Pb was produced from Balmat from 1927–1978 alone, and in 1980 production from this mine alone amounted to 14% of total zinc production in the United States (deLorraine and Dill 1982). The district is hosted in the Upper Marble sequence of deLorraine and Carl (1986), the stratigraphic equivalent of the Upper Marble and Quartzite package of Wynne-Edwards (1967a). Structural reworking has resulted in the concentration of ore in fold hinges in some instances, however, despite structural and metamorphic reworking, deLorraine and Dill (1982) considered the original deposits to be stratiform and premetamorphic in origin. Deposition probably occurred in a shallow-water environment, possibly a sabhka, as anhydrite has been reported from the Balmat Mine (deLorraine and Dill 1982), and stromatolitic dolomites are common throughout the Upper Marble unit both at the mine site (Easton, unpublished data) and elsewhere in the near mine stratigraphy (deLorraine and Carl 1986).

It is problematic if such a deposit exists in Ontario. The facies variation in the Frontenac Terrane indicates a deepening basin to the west and northwest, and calcitic marbles predominate in the terrane in Ontario. Geochemically, regional till and lake sediment surveys (*see* Geochemistry) do not show high background levels of lead, zinc, or cadmium within the Frontenac Terrane. Improved knowledge of the stratigraphy and facies of the carbonate rocks within the Frontenac Terrane is needed before the stratiform zinc potential of the Frontenac Terrane in Ontario can be properly evaluated.

## IRON-TITANIUM DEPOSITS

Three ilmenite-ilmenomagnetite deposits are known from the Frontenac Terrane, all are associated with gabbroic or gabbroic anorthositic bodies. These are the Allen, Crosby Lake and Newboro deposits (Carter 1984). The Newboro deposit, also known as the Matthews–Chaffey deposit, is by far the largest and achieved limited production in 1853–1859 and 1870–1871 (Carter 1984). Carter (1984) reports reserves of 45 million tons, grading 26.2% Fe and 6.6% $TiO_2$, but containing about 1% S. Potential for other such deposits in the area is restricted to the northwestern part of the Frontenac Terrane, in association with other gabbroic anorthosite bodies such as the Tichbourne, Victoria and Blessington.

## SULPHIDE DEPOSITS

Carter (1984) reports 4 small pyrite deposits hosted in rusty quartzofeldspathic gneisses from the Frontenac Terrane, namely, the Brockville Chemical, Shipman and Sloan deposits. These would appear to be concentrations of syndepositional pyrite during metamorphism. All lie near the Paleozoic unconformity, and possible concentration during weathering prior to Paleozoic deposition (*see* Post-Grenville Events) may have been a factor in the genesis of these deposits.

## IRON SKARN DEPOSITS

These deposits are much less common in the Frontenac Terrane than in the Bancroft–Elzevir terranes, despite the abundance of carbonate rocks in the Frontenac Terrane. Carter (1984) reports only 3 deposits, all located in the northeastern part of the Frontenac Terrane near Perth. Two are associated with the margin of the Christie Lake gabbro, with the third being related to a monzonite suite pluton. The paucity of such deposits in the Frontenac Terrane may relate to, the fact that regional metamorphism and granitic plutonism were largely contemporaneous, or alternatively, the chemical composition of the host carbonates. The absence of volcanic rocks in the region may have meant a limited source of iron was available to be incorporated in the metasedimentary rocks.

## GRAPHITE

Several graphite properties are present in the northwest and west parts of the Frontenac Terrane, all within about 10 km of the Rideau–Canoe Lake fault. Many were producers at the turn of the century, but recently many properties have been re-evaluated (*see* Papertzian and Kingston 1982; Black and MacKinnon 1988; MacKinnon and LeBaron 1990). The 6 largest properties are the Globe Mine, the Cornell property, the Timmins Mine, the Burridge marble belt, the Desert Lake property and the Bawden Mine. The Globe and Cornell deposits have an acicular flake, whereas the other 4 properties all show equidimensional flakes (Black and MacKinnon 1988). Most occurrences have been known since the first part of the century, but only recently have they been actively explored and evaluated. The following material is largely summarized from Black and MacKinnon (1988).

The Globe Mine operated between 1870 and 1919 and was the first graphite property mined in Ontario. Reported reserves are estimated at 45 400 t of average 7% graphite. The Timmins Mine has reserves of 227 000 t of 8% graphite, although higher tonnages are likely present. This deposit is hosted in a narrow septum between the Westport and Rideau Lake monzonites, and contact metamorphism by these intrusions may have been a factor in graphite mineralization. The Cornell deposit has a grade of 10% graphite, possibly 20% locally, and is still undergoing evaluation. The Desert Lake property occurs only 150 m from the Rideau–Canoe Lake fault, and marble breccia zones are located on the property. Reserves of 4.5 million t of 7.6% graphite have been reported on the property. The Bawden Mine can be considered an extension of the Desert Lake property. Locally, it contains 20% graphite, with overall grade of 5 to 8%.

Almost all the deposits are hosted in calcitic marbles containing quartzose and rusty quartzose gneiss bands, somewhat similar to the "quartzite-limestone" transition rocks of Wynne-Edwards (1967a). Many are present in tightly folded synforms; all have been strongly deformed.

### BUILDING STONE

Several quarries for dimensional stone operate within Frontenac-type intrusions, specifically the Battersea and Lyndhurst granites (LeBaron et al. 1989). LeBaron et al. (1989, 1990) summarize the building stone potential of eastern Ontario, including the Frontenac Terrane.

### OTHER INDUSTRIAL MINERALS

Wollastonite occurs as a contact metamorphic phase to the Seeley's Bay gabbro (Wynne-Edwards 1962) and in the contact zone of the Lyndhurst Granite near Deans Island (Wynne-Edwards 1967a). MacKinnon (1990) summarizes the major wollastonite occurrences in the Frontenac Terrane. Calcitic marbles are present throughout the Frontenac Terrane, but most outcrops contain impurities, limiting their use in the filler industry (*see also* Grant et al. 1989).

Garnet and sillimanite are abundant throughout much of the terrane in the quartzofeldspathic gneisses, but economic concentrations have not been reported. Most garnet is andradite (Wynne-Edwards 1967a). At the turn of the century, mica and feldspar were mined actively from many sites in the Frontenac Terrane (*see also* listing *in* Hewitt 1964a), all of which remain potential sources of these 2 commodities.

Vermiculite occurs near Stanleyville in diopside-bearing silicate gneisses associated with marbles. Weathering of the gneisses during the Paleozoic may have played a role in forming these deposits, and Paleozoic rocks crop out near the occurrences. MacKinnon et al. (1990) describe the geology of the Stanleyville occurrences, as well as the industrial applications of wollastonite and other vermiculite prospects in southern Ontario.

# POST-GRENVILLE EVENTS

## Mafic Intrusions

Several small mafic to ultramafic intrusions are present in the North Bay and Burwash areas, generally near the southern margin of the Ottawa–Bonnechere Graben or near the southeastern margin of the Grenville Front Tectonic Zone. These include ultramafic stocks at Cleland Lake, Hagar Township, Barlow Lake and Trout Lake (Lumbers 1975) and gabbroic rocks at Caribou Lake (Friedman 1957), the Chebogomog Lake peridotite (Lumbers 1971a) and the Memesagamesing Lake Norite (Friedman 1955; Lumbers 1971a). All ultramafic bodies consist of partly serpentinized peridotite with phases of mafic-rich olivine gabbro (Lumbers 1975), some of which contain corona textures (Friedman 1955).

The age of these bodies is not well-constrained. Lumbers (1975) notes that they are postmetamorphic, and that one of the stocks is cut by a Grenville Swarm diabase dike. This would seem to give a minimum age of *circa* 575 Ma and a maximum age of *circa* 900 Ma.

Pyrite, pyrrhotite and chalcopyrite mineralization is associated with many of these intrusions (e.g., Caribou and Memesagamesing Lake bodies); recently, some have been examined for platinum mineralization (Garland et al. 1989, 1990).

## Mafic Dike Swarms

Diabase dikes are not as abundant in the Grenville Province as in other parts of the Canadian Shield. Dike swarms older than *circa* 1000 Ma have been metamorphosed and deformed by the Grenville Orogeny and are therefore harder to recognize as distinct swarms. An example is the 1235 million-year-old Sudbury dike swarm which straddles the Grenville Front into the Beaverstone Domain of the Grenville Province (Davidson and Bethune 1988).

Four dike sets are present within the Grenville Province as shown on Figure 19.92; namely: 1) basaltic dikes of the *circa* 900 million-year-old northwest-trending Kingston (Frontenac) Swarm; 2) basaltic dikes of the *circa* 590 to 550 million-year-old east-striking Grenville Swarm and 3) the *circa* 450 to 420 million-year-old northeast-trending Rideau Swarm; and 4) Jurassic alkalic and lamprophyric Picton dikes (Fahrig and West 1986). Although mapped (cf., Lumbers 1971a, 1975; Card and Lumbers 1977; Fahrig and West 1986), the dike rocks themselves have not received much study. Lumbers (1971a, 1975) provides the most complete descriptions. All 4 sets are associated with younger geologic structures, mainly the Ottawa–Bonnechere and St. Lawrence graben systems (*see* Figure 19.92). None has been studied in detail.

### GRENVILLE SWARM

Grenville Swarm dikes are discontinuous, occupying narrow, long, linear east-striking, fault-controlled valleys in a region roughly 75 km north and south of lat. 46°N and are generally vertical. The French, Pickerel and Seguin rivers occupy such valleys (Lumbers 1971a, 1975; McRoberts 1991).

Diabase dikes range in thickness from 50 cm to 30 m and can pinch and swell in thickness along strike. Lumbers (1971a, 1975) reports that the rocks are medium-grained, with subophitic texture, and locally contain coarse pegmatitic patches. Andesine labradorite, calcic clinopyroxene and pigeonite, with minor oxides and apatite, are the dominant mineralogy. Some dikes are sheared and brecciated by postemplacement faulting and contain disseminated sulphide mineralization (Lumbers 1971a).

Murthy (1971) refers to the dikes as quartz diabase on the basis of their mineral constitution. Kretz et al. (1985) studied several of the dikes in the Hull area and provided several chemical analyses confirming their tholeiitic character.

Some workers had suggested a Neoproterozoic age for the dikes (e.g., Lumbers 1971a, 1975; Card and Lumbers 1977), and Kretz et al. (1985) suggest an age of *circa* 700 Ma. Kretz et al. (1985) also suggest that there may be 2 ages for the dikes in the Hull area. York et al. (1991) report an Ar-Ar age of 570±3 Ma from a dike near Chalk River.

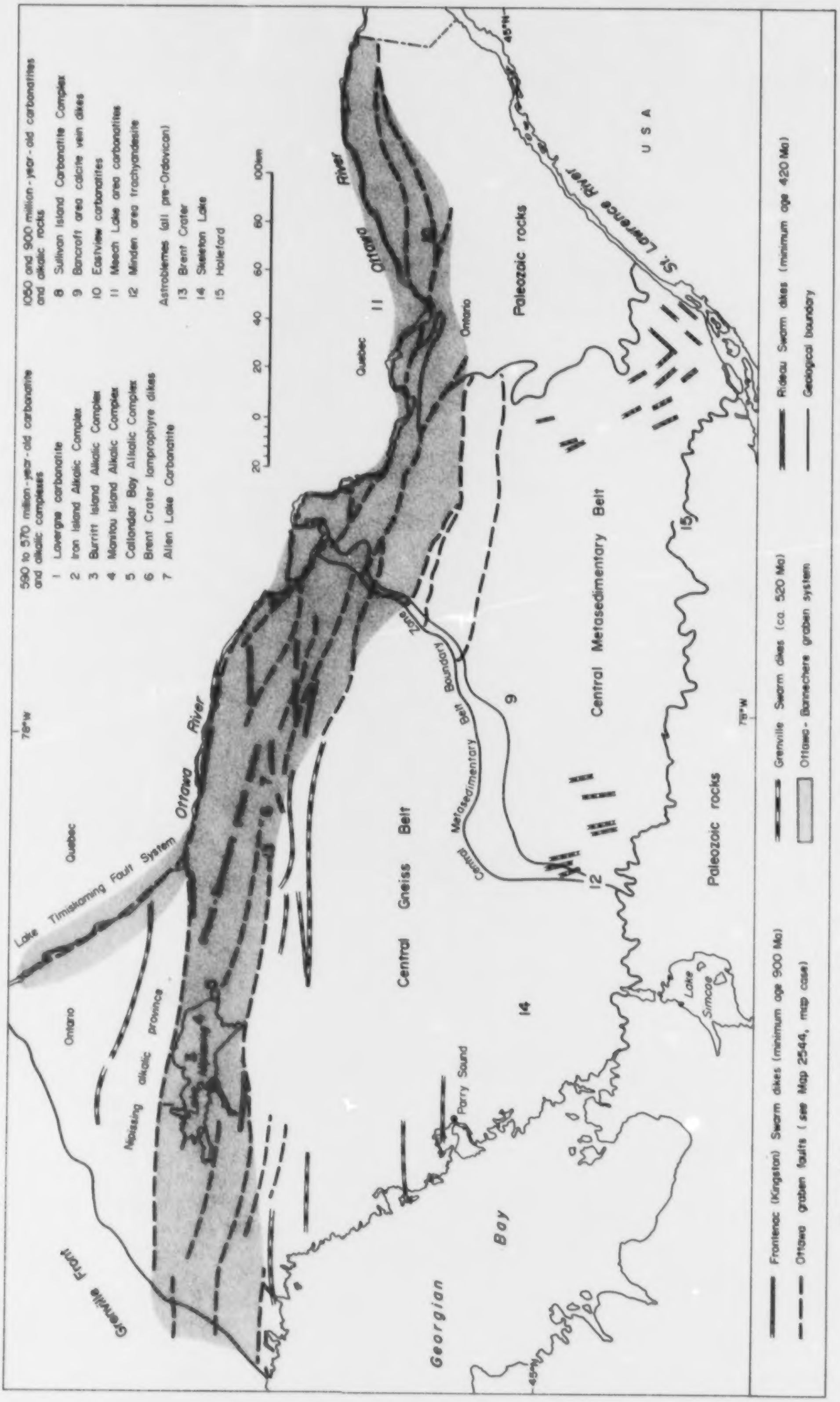

590 to 570 million-year-old carbonatite and alkalic complexes
1 Lavergne carbonatite
2 Iron Island Alkalic Complex
3 Burritt Island Alkalic Complex
4 Manitou Island Alkalic Complex
5 Callander Bay Alkalic Complex
6 Brent Crater lamprophyre dikes
7 Allen Lake Carbonatite
1050 and 900 million-year-old carbonatites and alkalic rocks
8 Sullivan Island Carbonatite Complex
9 Bancroft area calcite vein dikes
10 Eastview carbonatites
11 Meech Lake area carbonatites
12 Minden area trachyandesite
Astroblemes (all pre-Ordovician)
13 Brent Crater
14 Skeleton Lake
15 Holleford
20 0 20 40 60 80 100km
Frontenac (Kingston) Swarm dikes (minimum age 900 Ma)
Ottawa graben faults (see Map 2544, map case)
Grenville Swarm dikes (ca. 520 Ma)
Ottawa-Bonnechere graben system
Rideau Swarm dikes (minimum age 420 Ma)
Geological boundary
Grenville Front
Georgian Bay
Parry Sound
Nipissing alkalic province
Ontario
Quebec
Lake Timiskaming Fault System
Ottawa River
Central Gneiss Belt
Central Metasedimentary Belt Boundary Zone
Central Metasedimentary Belt
Paleozoic rocks
Lake Simcoe
St. Lawrence River
U S A
78°W
75°W
45°N
9
11
12
14
15

Figure 19.92. Distribution of post-Grenville Orogeny dikes, the Ottawa–Bonnechere graben system and Neoproterozoic–Early Cambrian alkalic complexes.

Kumarapeli (1985) has suggested that the Grenville dike swarm is related to the Tibbit Hill volcanics in Quebec, and that the Ottawa–Bonnechere Graben represents a failed rift with the Tibbit Hill volcanics being deposited at the apex of the rift. The Tibbit Hill volcanics have been dated by U-Pb zircon at $554^{+4}_{-2}$ Ma (Kumarapeli et al. 1989).

### KINGSTON (FRONTENAC) SWARM

Dikes related to this northwesterly to northerly trending, shallowly dipping swarm are most abundant in the Frontenac Terrane near Kingston, however, other dikes possibly correlative with this swarm occur throughout the Central Metasedimentary Belt from Minden to Sharbot Lake (Wolff 1985; Bright 1987, 1988; Easton 1987e, 1990a). The age of this swarm is poorly constrained; K-Ar whole-rock ages between 761±75 Ma and 860±30 Ma (Wanless et al. 1967, 1974) have been reported from dikes of this swarm near Kingston. Easton and Roddick (1988) report a K-Ar whole-rock age of 902±12 Ma from a weakly metamorphosed diabase dike near Minden possibly correlative with the Kingston Swarm. A. Davidson (Geological Survey of Canada, personnel communication, 1992) has suggested that the dikes may be older than 900 Ma in age. Geochemical analyses of the dated dike, as well as others of similar relative age in the Minden area, indicate an alkali basalt composition (Easton 1987e, 1990a), but lie near the tholeiite-alkalic basalt transition. In thin section, the dikes contain equal proportions of chlorite, hornblende and relict clinopyroxene and andesine. Some dikes are weakly foliated, others are more strongly sheared and all appear to have been subjected to low-grade metamorphism or alteration, consistent with their probable *circa* 900 million year age.

### RIDEAU SWARM

These northeast-trending, vertical mafic dikes have only been reported from the Kingston–Gananoque area. Park and Irving (1972) report 2 K-Ar whole-rock ages of 416±23 Ma and 420±23 Ma from dikes of this swarm, and another dike was dated at 442±75 Ma (Wanless et al. 1966). However, Wright (1923) described the northeast-trending dikes of the Mallorytown area as being older than the Paleozoic cover rocks, which would suggest a minimum age of roughly 475 to 525 Ma (Harland et al. 1989). In thin section, unweathered dikes are ophitic-textured, slightly porphyritic and composed of labradorite, augite and iron oxide.

Andesine porphyritic and amygdular andesite dikes cut the diabases in the southeast part of the Westport area and in the Gananoque area (Wynne-Edwards 1962, 1967a).

## Fault Systems

### OTTAWA-BONNECHERE GRABEN SYSTEM

The Ottawa-Bonnechere graben system runs from Montréal to Lake Timiskaming (*see* Figure 19.92) and controls the course of the Ottawa River. The Ottawa graben displays a rift valley morphology and is about 60 km wide and 700 km long (Kumarapeli 1985). North of Ottawa, a major splay, the Bonnechere part of the system, separates from the main trend. It is an east-trending zone from Renfrew to North Bay to the north shore of Lake Huron, roughly coincident with lat. 46°N (Kay 1942; Kumarapeli and Saull 1966; Kumarapeli 1976, 1985). Mereu et al. (1986a, 1986b) describe the results of a seismic refraction survey across the Ottawa graben, which shows an ill-defined Moho underneath the graben and major faults corresponding with the graben margins (*see* Figure 19.10c). Although the Ottawa graben is discordant to Grenville structural trends, Kumarapeli (1985) notes that the graben faults are controlled by a regional fracture pattern within the shield, which reflects long-lived crustal structures. The southern boundary of the Ottawa graben in the Mattawa–North Bay–French River area is roughly coincident with the boundary between Archean and Paleoproterozoic to Mesoproterozoic crust within the Grenville Province that Dickin and McNutt (1989a) suggest represents a Penokean suture zone.

The fault system developed sometime after the peak of Grenville metamorphism and has been active intermittently since that time, with nodes of activity at *circa* 575 Ma (mafic dike injection [Grenville Swarm], Tibbit Hill volcanics and alkalic magmatism), *circa* 450 to 420 Ma (faulting resulting in eventual preservation of Paleozoic outliers across the Grenville Province, local basaltic magmatism [Rideau Swarm]) and *circa* 190 to 170 Ma (ultrapotassic and ultramafic magmatism [Picton dikes]). Recently, earthquake activity has occurred locally along the Ottawa–Bonnechere graben system (Adams and Basham 1989). Kumarapeli (1985) regards the main development of the Ottawa–Bonnechere graben system as Early Paleozoic to Neoproterozoic, and the graben system (rift valley) formed as a consequence of Iapetan (Middle Paleozoic) rifting. Reactivation has occurred locally along the essentially Early Paleozoic feature.

### ST. LAWRENCE GRABEN SYSTEM

The St. Lawrence graben, in contrast, developed in the Mesozoic, coincident with development of the present Atlantic Ocean (Kumarapeli 1985). However, development of this younger graben system, in part, reactivated the Ottawa–Bonnechere graben system (Kumarapeli and Saull 1966; Kumarapeli 1985). The system is well-defined from the mouth of the St. Lawrence to Montreal, but its western extent is unknown. Adams and Basham (1989) show that it extends beyond Kingston into Lake Ontario and Lake Erie, mainly on the basis of recent seismicity. Some major aeromagnetic features, such as the southward extension of the Central Metasedimentary Belt Boundary Zone, appear to be offset along a linear feature running under Lake Ontario that may be part of this system. A variety of neotectonic features in the Lake Ontario region have also been attributed to the extension of this system (*see* Johnson et al., this volume).

### MINERALIZATION

Narrow, small, post-Ordovician barite-fluorite-strontium-lead-rich veins occur in Grenville marbles along the Ottawa

River and near the Precambrian-Paleozoic unconformity, primarily in the Madoc area. Most reviews of Grenville mineral deposits include these veins with other Grenville mineral deposits (e.g., Sangster 1970; Sangster and Bourne 1982; Carter 1984), even though they are clearly post-Grenville in age and genetically related to the Ottawa–Bonnechere graben system and the Precambrian- Paleozoic unconformity (Carter 1984).

Most veins are localized along vertical to near-vertical west and northwest-striking fractures that parallel faults of the Ottawa–Bonnechere Graben or north-northwest-striking faults. The majority are hosted in Grenville marbles or Paleozoic limestone, near the Precambrian-Paleozoic unconformity. Some are hosted in Grenville supracrustal and intrusive rocks. Coarse-grained, white to pink calcite is the principal vein mineral, with extremely varied proportions of barite, fluorite, galena and locally sphalerite and celestite (Carter 1984). According to Sangster (1970) and Sangster and Bourne (1982), the 3 most common vein types are calcite-fluorite-barite, calcite-barite-galena and calcite- celestite. Further details on the metallogeny of these veins is given in Easton and Fyon (this volume).

Fluorite-bearing veins are most common in the Madoc area, where they have been described by Guillet (1964). Easton (1989a) drew attention to the fact that the veins are closely associated with the Deloro Granite, a unique high-fluorine, alkalic granite that has been subjected to a *circa* 450 million-year-old low temperature, isotopic-resetting event (Lopez-Martinez and York 1983; Hanes et al. 1988). It appears that the source of the fluorine in the veins is related to the Deloro Granite, accounting for their concentration in the Madoc area.

Calcite-barite-galena veins are more common and occur over a wider area, being particularly abundant in the Renfrew–Arnprior area. The Kingdon Mine (Carter et al. 1980; Easton et al. 1986) is a past producer. Calcite-celestite veins occur mainly in the Kingston–Sharbot Lake area underlain by the Frontenac Terrane. As in the case of the Madoc veins, the presence of celestite may reflect the chemical character of the country rocks (i.e., the presence of a strontium-rich host, most likely syenite plutons of the Frontenac type).

Low-grade isotopic resetting of argon-argon and rubidium-strontium systems has been documented adjacent to the Precambrian-Paleozoic unconformity in the Havelock–Tweed region (Wanless and Loveridge 1972; Lopez-Martinez and York 1983; Dunlop and Stirling 1985; Hanes et al. 1988). This resetting is probably directly related to the formation of the calcite-lead vein systems in the region.

## Alkalic and Carbonatitic Intrusive Rocks

### NEOPROTEROZOIC ALKALIC MAGMATISM

Elsewhere in Ontario, Sage (this volume) reports several carbonatite and alkalic complexes emplaced within the time span 1250 to 900 Ma. As discussed in the section under the Central Metasedimentary Belt, a period of carbonatitic and alkalic magmatism probably occurred in the Grenville Province between 1070 and 1030 Ma (Lumbers et al. 1990). However, there may have also been a post-Grenville period of carbonatitic and alkalic magmatism in the period 950 to 850 Ma.

Easton (1986b, 1990a) and Easton and Roddick (1988) describe a trachyandesite dike along the Central Metasedimentary Belt Boundary Zone near Moore Falls, south of Minden, that has a K-Ar biotite age of 898±17 Ma (Easton and Roddick 1988). The trachyandesite dike is chemically similar to a trachyandesite flow described by Lafleur and Hogarth (1981) from Buckingham, Quebec (Easton 1986b, 1990a) dated by K-Ar at 573±32 Ma (Lafleur and Hogarth 1981), similar to ages obtained on rocks from the Nipissing alkalic province (Currie 1976). Perhaps intermittent alkalic magmatism occurred along major crustal features in the Grenville Province throughout the Neoproterozoic to Early Paleozoic.

Lumbers (1982) describes possible Neoproterozoic to Early Paleozoic age trachyte dikes near Calabogie Lake and along the Barron River. These may be part of this alkalic suite.

Hogarth and Rushforth (1986) and Hogarth (1966) describe carbonatite dikes in McCloskey's Field in the Hull area that have been dated at 930±10 and 910±10 Ma by K-Ar on phlogopite. Fenite from the nearby Haycock fenite, in Quebec, furnishes K-Ar phlogopite ages of 980±52 and 862±31 Ma (Hogarth and Rushforth 1986). Also in the Hull area, carbonatite dikes from McGerrigle's Field yield a K-Ar phlogopite age of 920 Ma (Currie 1976). All 3 areas suggest a period of post-Grenville, Neoproterozoic alkalic mag-matism, perhaps contemporaneous with that found in the Minden area.

### NEOPROTEROZOIC TO EARLY PALEOZOIC ALKALIC MAGMATISM

Sage (this volume) briefly noted the occurrence of alkalic and carbonatite complexes along the North Bay splay of the Ottawa–Bonnechere Graben, and elsewhere along the Ottawa graben, particularly where it intersects the St. Lawrence graben. Sage (this volume), however, did not provide details on the individual complexes found along the Ottawa–Bonnechere and St. Lawrence graben systems.

Currie (1976) describes the North Bay splay as the Nipissing alkalic province, containing 4 nearly circular central complexes 3 km in diameter spaced at intervals of 10 to 15 km (*see* Figure 19.92). Currie (1976) synthesizes the data on the Nipissing alkalic complexes, apart from the Allen Lake Carbonatite, discovered in 1977 by an airborne gamma-ray survey (Ford et al. 1988).

Extensive aureoles of fenitization and dikes occur around each complex, and nephelinitic lamprophyre dikes occur sporadically throughout the North Bay splay. The central complexes are 1) the Callander Bay Alkalic Complex (Currie and Ferguson 1971, 1972; Ferguson and Currie 1971, 1972; Lumbers 1971a), 2) the Manitou Island

Alkalic Complex (Rowe 1958; Lumbers 1971), 3) the Burritt Island Alkalic Complex (Lumbers 1971a), 4) the Iron Island Alkalic Complex (Lumbers 1971a) and 5) the Allen Lake Carbonatite (Ford et al. 1988). Many of these complexes are poorly exposed, and the centres of most are below the level of Lake Nipissing. Currie (1976) and Lumbers (1971a) also report 3 zones of fenitization, which may be related to subsurface alkalic rocks: 1) in Springer Township, also known as the Lavergne carbonatite (Sage, this volume); 2) on Highway 11, 6 km south of Powassan; and 3) east of Lake Nosbonsing near Deux Rivières along the Ottawa River. Brent Crater (Currie 1971, 1976) has also been interpreted as an astrobleme (Millman et al. 1960; Dence et al. 1971).

Lamprophyric dikes in the vicinity of Brent Crater have been dated by K-Ar whole-rock at 576±40 Ma (Currie 1976; Shafiqullah et al. 1968). Biotites from lamprophyre and nepheline syenite from the Callandar Bay Alkalic Complex have been dated by K-Ar methods at 568±14 Ma (Ferguson and Currie 1972) and 575 Ma (Currie 1976), respectively. The fenite zone of the Manitou Island Alkalic Complex (Newman Island) has been dated by K-Ar biotite at 580±40 Ma (Gittins et al. 1967) and 569±50 Ma (Lowdon et al. 1963).

The Manitou Island Alkalic Complex contains 5 deposits of uranian pyrochlore; the largest deposit lies just east of Newman Island (Currie 1976). Reserves of 1 893 000 tons of 0.86% $Nb_2O_5$ and 2 962 000 tons of 0.041% $U_3O_8$ and 0.69% $Nb_2O_5$ have been reported from the Newman Island deposit (Currie 1976; Rowe 1958). No significant mineralization has been reported from the other complexes (Currie 1976), although Ford et al. (1988) note enrichment of iron, niobium, thorium and the REE in the Allen Lake Carbonatite. Ford et al. (1988) provide a detailed survey of the geochemical dispersal train found in till, lake sediment and vegetation in the vicinity of the Allen Lake Carbonatite, all of which could be used successfully for prospecting in the Nipissing alkalic province.

## MESOZOIC ALKALIC MAGMATISM

Hogarth and Rushforth (1986) describe the Eastview carbonatite dikes cutting Paleozoic strata, along the Ottawa graben near Blackburn, west of Ottawa. Hogarth and Rushforth report that the paleomagnetic pole of the dikes coincides with that of the Oka Complex in Quebec, suggesting a Cretaceous age.

Barnett et al. (1984) have described Jurassic micaceous kimberlite dikes near Belleville, near the western end of the St. Lawrence rift valley. The dikes are referred to as the Varty and Picton dikes and may be related to the Picton–Napanee fault, part of the Robertson Lake mylonite zone, and the Clarendon–Linden Fault. Barnett et al. (1984) note that the dikes have both kimberlitic and lamprophyric affinities. Sage (this volume) reports a U-Pb age of 176±3 Ma (Middle Jurassic) on the Varty dike, consistent with K-Ar ages of 171 Ma on the Varty dike and 175 and 173 Ma from the Picton dike (Barnett et al. 1984). McFall (1990) reports slickensides from the Picton dike, indicating syn- to post- Jurassic movement on the dike and presumably, the Picton–Napanee fault.

## Astroblemes

At least 3 astroblemes occur in the Grenville Province, the craters of which are filled with Paleozoic strata indicating that the impacts occurred sometime after uplift and erosion of the Grenville Province but before Paleozoic weathering. The 3 astroblemes are Brent Crater (Millman et al. 1960; Dence et al. 1971; Lozej and Beales 1975), located along the Ottawa–Bonnechere Graben between North Bay and the Ottawa River; Skeleton Lake, located in the Muskoka region between Rosseau and Huntsville (Waddington and Dence 1979); and Holleford Crater, located just south of the Precambrian-Paleozoic boundary northwest of Kingston (Beals 1960; Dawson 1961; Bunch and Cohen 1963; St. John 1968).

Brent Crater has also been regarded as an alkalic complex (Currie 1971), however, it lacks the intense magnetic and gamma-ray spectrometric signatures associated with other alkalic complexes along the Ottawa–Bonnechere Graben, and thus, an impact origin is more likely. In addition, K-Ar dating of fractured rocks from the vicinity of Brent Crater give a minimum age of 424±20 Ma (Hartung et al. 1971), although there is considerable excess argon in the fractured rocks of Brent Crater. This is younger than the 590 to 560 million-year-old K-Ar ages of alkalic complexes in the North Bay area and suggests that Brent Crater is unrelated to the Nipissing alkalic province (although fenites and lamprophyre dikes do occur in the vicinity of Brent Crater [Lumbers 1971a; Currie 1976]).

## Erosion and Uplift

Kay (1942) notes that the Grenville Province was eroded to form a peneplain (Cashel peneplain) prior to deposition of Paleozoic strata in southern and eastern Ontario. As noted in an earlier section on the uplift history of the Grenville Province, argon became trapped in biotite across the Grenville Province at about 930 Ma, indicating cooling below 300°C. Thus, uplift and erosion of several tens of kilometres of rock and peneplain formation took place over 400 million years. Di Prisco and Springer (1991) suggest that weathering prior to deposition of Middle Ordovician carbonates may have lasted at least 350 million years. If so, this would mean uplift and erosion following the Grenville Orogeny was mostly complete by 800 Ma.

## Pre-Paleozoic Weathering of the Shield and the Precambrian-Paleozoic Unconformity

### EVIDENCE OF PRE-PALEOZOIC WEATHERING OF THE PRECAMBRIAN SHIELD IN CENTRAL AND SOUTHERN ONTARIO

Di Prisco and Springer (1991) documented the existence of an exhumed Precambrian paleosurface beneath Ordovician carbonate rocks. In the Madoc–Marmora area, the surface is characterized by: 1) intensely weathered surfaces which are inconsistent with post-glacial weathering effects, 2) the presence of hematite above and below the unconformity and 3) karst features. Outcrop descriptions also indicate that there has been an increase in porosity of the Precambrian rocks within the zone affected by weathering.

Di Prisco and Springer (1991) describe weathering and minor mineralization associated with the unconformity. Mineralization includes: 1) hematization of Grenvillian magnetite skarn deposits to depths of several hundred metres below the unconformity; 2) concentration of minerals, particularly gold, in the regolith developed along the unconformity; 3) weathering of phlogopite-rich horizons to form vermiculite; 4) localized concentration of manganese and uranium along the unconformity; and 5) development of karst-systems where chalcopyrite and gold are locally deposited (e.g., Eldorado Mine, Ontario's first gold mine). Minor transport of mineralization has occurred along the unconformity, and most mineralized areas represent enrichments of pre-existing Grenvillian mineralization. Di Prisco and Springer (1991) also describe many of the sites where the Precambrian-Paleozoic unconformity is exposed in southern Ontario, as do Easton (1987h) and Easton et al. (1990).

## BASEMENT ALTERATION ZONE IN SOUTHWESTERN ONTARIO

Carter and Easton (1990), Easton and Carter (1990, 1991) and Harper et al. (1991) have documented the presence of a regional-scale alteration zone in the uppermost portion of the Precambrian basement rocks of southeastern Ontario. The altered zone extends, on average, 2 to 5 m below the Paleozoic-Precambrian unconformity. The most noticeable feature of this alteration zone is saussuritization of plagioclase. Mafic minerals commonly have corroded margins in the alteration zone and in some cases have been chloritized. Salmon red potassium feldspar is more common within the altered zone than in the fresh basement rocks and is presumed to be authigenic in origin (Harper et al. 1991). Similar features have been noted in Ohio by Mensing and Faure (1983) and by Duffin (1989) in the US midcontinent extending from Ohio to Iowa, Missouri and Wisconsin.

## NEOPROTEROZOIC TO EARLY PALEOZOIC SEDIMENTATION

Johnston et al. (this volume) describe the lower Paleozoic section in southern Ontario. However, locally a few thin, discontinuous, sedimentary units are present in southern Ontario and northeastern New York between the Precambrian surface and the lowermost Paleozoic units (Shadow Lake, Nepean and Gull River formations; Di Prisco and Springer 1991; Bloomer 1965, 1967; Brown 1967). In Ontario, these consist mainly of coarse gritstones to pebble conglomerates of the Covey Hill Formation. They are commonly hematized and are associated with impure arenites of local derivation which contain angular fragments of basement rocks (Wolf and Dalrymple 1984). Maximum thickness is 2.3 m (Wolf and Dalrymple 1984). If these rock units are older than the basal Paleozoic section, then they represent a discontinuous sedimentary unit of local derivation, probably deposited mainly on an exposed, subaerial surface. In addition, the Covey Hill Formation is unconformably overlain by the Nepean Formation and is geographically restricted to an area adjacent to the Rideau Lake fault in the Frontenac Terrane (Wolf and Dalrymple 1984). This coarse conglomerate and arkosic sequence may be Precambrian in age.

# Extension of the Grenville Province Beneath the Paleozoic of Southwestern and Southern Ontario

## INTRODUCTION

Due to its location on a structural arch and the consequent thinness of Paleozoic cover rocks, southwestern Ontario is a uniquely favourable location for the study of the Precambrian basement rocks in the eastern midcontinent of North America. Mesoproterozoic rocks of the Grenville Province outcrop to the northeast of Toronto and can be traced southwestward beneath the cover sequence utilizing drill-hole data and gravity and aeromagnetic maps. The close association of many oil and gas reservoirs in southwestern Ontario with faults has been documented by Sanford et al. (1985), particularly Ordovician and Cambrian reservoirs (*see* Johnston et al., this volume). Basement features may have had a controlling influence on the location of faults cutting the Paleozoic section, and an understanding of the basement geology is required in order to locate faults and associated oil and gas reservoirs.

Approximately 600 drill holes, mostly oil and gas wells, have penetrated the Grenvillian basement in southwestern Ontario (Figure 19.93). Carter and Easton (1990) and Easton and Carter (1990, 1991) used well data, in conjunction with interpretation of aeromagnetic and gravity maps, to document the distribution of basement rock types and identify several lithotectonic domains in southwestern Ontario. These lithotectonic domains are illustrated in Figure 19.94 and are of the same scale and dimensions as major terranes present within the Central Gneiss Belt exposed to the north. Harper et al. (1991) and Carter and Easton (1990) have also documented the extent and characteristic features of a zone of alteration associated with a weathered zone developed on the Precambrian surface at the Precambrian-Paleozoic unconformity.

## LITHOTECTONIC DOMAINS

Failure to recognize the alteration zone described above thwarted previous attempts to precisely map the basement rock types (Turek and Robinson 1982a, 1982b; Bright 1986b), since most of the basement drill holes do not penetrate this zone. Boundaries of most of the domains shown in Figure 19.94 are tentative. From southwest to northeast these are the extension of the Grenville Front and the Grenville Front Tectonic Zone (GFTZ), the Huron Domain, the

**Figure 19.93.** Drill core intersecting the Precambrian-Paleozoic unconformity in southwestern Ontario. Gneissic tonalite is visible to the left. Centre core shows regolithic material derived from Precambrian rocks. Overlying Paleozoic strata (mainly coarse sand) is visible in the centre part of the right core (photograph *courtesy of* T.R. Carter).

Kent Domain, the Waterloo Domain, the Alliston Domain, the Cambridge Domain, the extension of the Fishog Domain and the extension of the Central Metasedimentary Belt Boundary Zone.

The Grenville Front cuts south beneath the centre of Lake Huron and along the St. Clair and Detroit rivers. The Grenville Front Tectonic Zone, at least in Essex County, extends from the Grenville Front east to Point Pelee, where aeromagnetic trends and basement rocks change in character. This position of the Grenville Front is placed some 50 km farther east of where previous workers had placed it (McLaughlin 1954; Nwachukwu et al. 1965; Turek and Robinson 1982a, 1982b; Gordon and Hempton 1986; Lidiak and Hinze, in press) based on the recognition that the aeromagnetic anomalies associated with the Killarney Magmatic Belt do not mark the Grenville Front (Green et al. 1988; McGrath et al. 1988). Drill core studies in Ohio by Lucius and Von Frese (1988) and seismic data from Lake Erie (Forsyth et al. 1991) confirm this eastward shift in the position of the Grenville Front shown in Figure 19.94.

The GFTZ in Essex County in southwestern Ontario has a pronounced westerly to southwesterly linear trend of its associated aeromagnetic pattern, coincident with, and on the extension of, a gravity high which marks the location of the Midcontinent Rift (MCR) in Michigan. The GFTZ also coincides with a west to northwest-trending belt of tonalitic rocks in the basement of Essex County. Perhaps these tonalites are metamorphosed equivalents of Keweenawan clastic rocks which were deposited as a rift fill sequence within the Midcontinent Rift in Michigan (Dickas 1986).

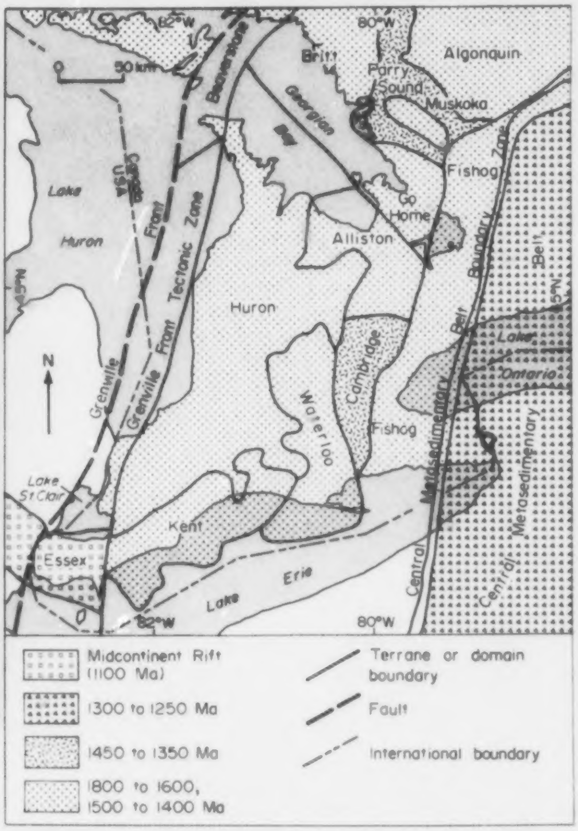

82°W
80°W
0
50 km
Beaverstone
Britt
Parry Sound
Algonquin
Muskoka
Georgian Bay
Fishog
Go Home
Alliston
Lake Huron
Canada USA
Front
Grenville Front Tectonic Zone
45°N
Huron
N
Cambridge
Waterloo
Fishog
Lake St Clair
Kent
Essex
Lake Erie
Grenville
Central Metasedimentary Belt Boundary Zone
Lake Ontario
Central Metasedimentary Belt
43°N
Midcontinent Rift (1100 Ma)
1300 to 1250 Ma
1450 to 1350 Ma
1800 to 1600, 1500 to 1400 Ma
Terrane or domain boundary
Fault
International boundary

**Figure 19.94.** Lithotectonic domains in the Precambrian basement of southwestern Ontario interpreted from drill core and cuttings and published aeromagnetic maps. Patterns refer to inferred crustal ages. *Modified from* Carter and Easton (1990).

Proposed extensions of the boundaries of the Britt and Go Home domains along Georgian Bay are shown in Figure 19.94, based on extension of the strong northeast-trending aeromagnetic pattern of these domains under the northeastern half of Georgian Bay.

The Huron Domain is characterized by a broad, open pattern of low intensity and relief on aeromagnetic maps. Lithologically it is dominated by gneissic quartzofeldspathic rocks ranging from granitic to tonalitic composition. Rare gabbroic and metasedimentary rocks occur.

The Kent, Waterloo and Alliston domains are correlative with areas of relatively high magnetic intensity. Granitic rocks are predominant in all 3 domains, with an abundance of granodioritic to tonalitic rocks occurring in the Waterloo and Kent domains. In the Waterloo Domain, the magnetic structure has a preferred northerly trend, with several elliptical magnetic highs which are interpreted to be magnetite-bearing plutons.

The Cambridge Domain is the most lithologically distinctive in the subsurface of southwestern Ontario, containing a belt of northerly trending dolomitic marble and associated metaclastic rocks which correspond to an area of low aeromagnetic intensity and relief. The northerly orientation and presence of minor amounts of marble suggests that this domain may be analogous to the Parry Sound Domain. However, the low aeromagnetic relief and lack of mafic rocks are in direct contrast.

The Fishog Domain is an extension of the highly magnetic granite-tonalite-monzonite rocks exposed in the Central Gneiss Belt west of Minden and is aeromagnetically distinct. It appears to be terminated at the north shore of Lake Erie by a zone of aeromagnetic lows.

The Central Metasedimentary Belt Boundary Zone changes direction sharply, just north of Minden, paralleling and coinciding with a similar change in orientation of the Grenville Front. In contrast with the exposed portion of the Central Metasedimentary Belt Boundary Zone, its interpreted extension in southwestern Ontario is characterized by a narrow, strongly linear zone with a high magnetic intensity, which strikes south-southwest from Minden through the Niagara Peninsula just west of St. Catharines. This coincides with a previous interpretation for the Niagara Peninsula portion of the zone (Carter 1987).

Most of the wells east of the interpreted position of the Central Metasedimentary Belt Boundary Zone intersected metaclastic rocks analogous to metamorphosed sandstones, siltstones and mudstones that are common in the western portion of the Central Metasedimentary Belt. Diagnostic features include high proportions of quartz and/or mica, the presence of graphite and rock fragments with identifiable detrital grains. Grain size is also typically finer than for rocks to the west of the Central Metasedimentary Belt Boundary Zone.

## BASEMENT INLIERS IN SOUTHEASTERN ONTARIO

Exposures of Grenville basement occur in a few areas of southeastern Ontario, mainly along major rivers, or post-Grenville faults or both. Kay (1942) recognized 2 types of inliers: 1) areas of Precambrian rock in stream valleys, most commonly near the Shield-Paleozoic contact (e.g., Newburgh and Lonsdale inliers); and 2) topographic highs. The best exposed of these are located near Shannonville and Ameliasburg. Other examples are located near Naphan and the Lynch Rock, Galesburg, Red Rock and Rohalion inliers (Kay 1942). A variety of rock types are exposed in these inliers, including silicated marbles, granitic gneisses and, near Shannonville, conglomerates alternatively considered as Hastings Series (i.e., Flinton Group; Kay 1942) or basal Paleozoic rocks (Tovell and Freeman 1978). Several inliers along the trace of the Salmon River fault may be indirectly related to this structure.

# TECTONIC SUMMARY

## Overview of Tectonic Models for the Grenville Province in Ontario

Tectonic models proposed to explain the development of the Grenville Province in Ontario are summarized in Table 19.18. They fall into 2 broad categories. First are non-plate-tectonic models, which tend to emphasize the unique aspects of Grenville Province geology and which are largely applicable exclusively to the Grenville Province. Second are plate-tectonic models, mostly focussing on the development of part or all of the Central Metasedimentary Belt. Many models in both groups were designed to explain certain features of Grenville Province geology, such as AMCG magmatism or anorthosite genesis in particular, and thus are not specific to, and are sometimes at odds with, Ontario Grenvillian geology in general. Brief comments on these models are listed in Table 19.18. Two points are worth noting: 1) Many models have merit in explaining geologic features in the area for which they were specifically developed (e.g., central Elzevir Terrane) but are not broad enough to take into account interrelationships with the Central Gneiss Belt or other Central Metasedimentary Belt terranes. So far, no model has tried to outline the geologic development of both the Central Gneiss Belt and the Central Metasedimentary Belt for the period 1800 to 1000 Ma. 2) Many key geologic events in the Central Metasedimentary and Gneiss belts have yet to be dated, and, until such time as more comprehensive geochronologic data are acquired, any tectonic model is likely to be quickly superseded or contradicted by new geochronologic information.

## Concepts of Orogeny in the Grenville Province

With the onset of regional K-Ar geochronology studies across the Canadian Shield, it became apparent that the Grenville Province was characterized by *circa* 1000 to 950 million-year-old K-Ar ages on biotites (Cumming et al. 1955; Stockwell 1964). The K-Ar ages were viewed as cooling ages following a metamorphic episode that peaked at *circa* 1000 Ma. This metamorphic event was believed to be coincident with the "Grenville Orogeny", and the 1.0 billion year age has become synonymous in the literature with the "Grenville Orogeny."

It was initially believed that most structural and metamorphic features observed across the Grenville Province were the result of the *circa* 1.0 billion-year-old Grenville Orogeny. With the advent of Rb-Sr and U-Pb geochronologic studies in the Grenville Province, however, it became clear that older rocks were present within the Grenville Orogen. This led some workers, such as Wynne-Edwards (1967b, 1969, 1972), to suggest that much of the Grenville Province consisted of reworked Archean and Early Proterozoic crust, and that little juvenile crust existed in the region. It also became clear that not all structural and metamorphic events in the Grenville Province could be easily attributed to a single, 1.0 billion-year-old orogenic event. In part, this was because any orogenic event is by its very nature diachronous (*see also* Gilluly 1966). This led to various reinterpretations of the "Grenville Orogeny" (e.g., Wynne-Edwards 1967c; Stockwell 1982).

Moore and Thompson (1980), Baer (1981b) and Brock and Moore (1983), on the basis of stratigraphic and geochronologic studies, suggested that 2 orogenies were present within the Central Metasedimentary Belt, an older (ca. 1240 to 1150 Ma) "Elzevirian Orogeny" and a younger (ca. 1050 Ma) "Ottawan Orogeny", the latter being equivalent to the "Grenville Orogeny." All workers have regarded the younger event as being the more extensive event, as well as that responsible for regional metamorphism. Lumbers et al. (1990) also describe 2 orogenies and a late thermal and metasomatic event, however, they regard the older orogeny (Elzevirian) as being the more significant of the 2 and the event responsible for regional metamorphism. Thus, in their view, the "Elzevirian Orogeny" would correspond to the "Grenville Orogeny" (*sensu lato*). This chapter presents the view that the "Elzevirian Orogeny" represents a significant, and perhaps the major, orogenic event in the tectonic development of the Grenville Province in Ontario.

There is difficulty in correlating structural events between domains and terranes; deformation stages in orogenic zones may be locally diachronous, it is premature to introduce terms for additional orogenies.

## Tectonic Summary

Much of the history of the various terranes of the Grenville Province is summarized in Figures 19.95 and 19.96 which are summary time-space charts for the Central Gneiss Belt and the Central Metasedimentary Belt, respectively. In these diagrams, concurrent events in the various terranes can be readily seen, which help to elucidate the geologic and tectonic histories among the multitude of terranes present in both the Central Gneiss Belt and the Central Metasedimentary Belt.

Figure 19.97 is a cladogram outlining how the disparate pieces of the Grenville Province in Ontario may have been assembled. The Grenville Orogen is a complex thrust stack assembled over a considerable time interval. Terranes now adjacent to one another may have been separated by great distances in the past. Figure 19.97 serves to illustrate the complex nature of the Grenville Province in Ontario, and the difficulty in developing comprehensive models.

### PRE-1780 MILLION-YEAR-OLD ACTIVITY

Rocks older than 1780 Ma, restricted to Segments 2 and 3 of the Grenville Front Tectonic Zone and the Nipissing Terrane, are metamorphosed equivalents of Southern and Superior Province rocks. The age of "Grenville" metamorphism of these rocks is unknown, although K-Ar ages in the region are in the 1000 to 900 million year range and U-Pb zircon ages of pegmatites in the area, including lower intercept ages, lie between 1000 and 900 Ma. The intensity of "Grenville" metamorphism in this region is not clearly known. There is no evidence of any events unique to the Grenville Province between 2650 and 1780 Ma.

### 1780 TO 1680 MILLION-YEAR-OLD ACTIVITY

A major period of crustal development in the Central Gneiss Belt, particularly in the Algonquin Terrane, is indicated by Nd model ages (Dickin and McNutt 1989a, 1989b, 1990) from metasedimentary gneisses in the Algonquin Terrane. Younger plutonic rocks contain *circa* 1.8 billion-year-old model ages, suggesting derivation from mantle or lower crust of this age.

In the Algonquin Terrane, this interval is first marked by a period of sedimentation (ca. 1780 to 1700 Ma) on a basement of unknown character, but possibly of Penokean age. In the Tomiko Terrane, metasedimentary rocks were deposited at roughly 1687 Ma (Krogh 1989), perhaps a continuation of the *circa* 1780 to 1700 million-year-old sedimentation event seen in the Algonquin Terrane. In the southern Nipissing Terrane, supracrustal rocks of the French River area were deposited at *circa* 1744±11 Ma (Krogh 1989), possibly on Archean or Paleoproterozoic basement. In the Tomiko Terrane, there is no indication of Archean-age zircons within the metasedimentary rocks (Krogh 1989), indicating that the Tomiko depositional setting was isolated from Archean source regions. This is in contrast to the French River metasedimentary rocks, which do contain Archean-age zircons (Krogh 1989), indicating derivation, in part, from an Archean craton. Thus, although the Nipissing Terrane may be parautochthonous, the Algonquin and Tomiko terranes may have been allochthonous to the Superior Craton during their initial development.

In the Killarney area, volcanism and related magmatism occurred at *circa* 1740 Ma and may represent a continental arc sequence. In the Nipissing Terrane, the Sturgeon and West Bay granites may have been emplaced at *circa* 1740 Ma (Lumbers and Vertolli 1991), perhaps coincident with Killarney magmatism. In the Algonquin Terrane, sedimentation was followed by a period of granite intrusion and migmatization (ca. 1680 Ma, Corrigan 1990, Culshaw et al. 1991), the latter probably associated with regional metamorphism. The result of this *circa* 1680 million-year-old orogenic event (Easton 1992) was to form a broad region of juvenile continental crust, which was further stabilized during the subsequent 1450 to 1420 million-year-old magmatic event.

Nd-model ages on paragneisses (Marcantonio et al. 1990) hint that sedimentation occurred in the Frontenac Terrane during this interval. It is possible that the Frontenac Terrane represents a sliver of Central Gneiss Belt-type crust that may have been rifted from and then accreted to the North American continent during the Elzevirian Orogeny.

**Table 19.18.** Summary of tectonic models proposed to explain Grenville Province evolution in Ontario.

| | **Plate-Tectonic Models** |
|---|---|
| | **Grenville as a Whole** |
| Hoffman (1991) | **Windward Side of a Compressional Orogen:** Compressional orogen creates thick thrust stack, with the Grenville Front Tectonic Zone representing crustal-scale thrust ramp. Deep level of exposure and lack of foreland basin reflects high rates of rainfall over Grenville Orogen (Central Gneiss Belt). Carbonates and evaporates in Central Metasedimentary Belt deposited in rain shadow of orogen. Amazon-like drainage system carries eroded detritus into Arctic basins. |
| Hoffman (1992) | **Central Metasedimentary Belt as a Klippe:** CMB tectonically emplaced over older crust, rooted to the southeast beneath the Appalachains. |
| Emslie and Hunt (1989, 1990) | **AMCG (anorthosite-mangerite-charnockite-granite) Magmatism in a Plate-Tectonic Context:** Grenville Orogeny pre-1200 Ma, Grenville regional metamorphic event (ca. 1070 Ma) related to major anorogenic magmatism peaking at 1160–1130 Ma followed by granulite-facies metamorphism, with tectonism and mountain building related to subsequent continent-continent collision. Keweenawan rifting is extensional and related to AMCG magmatism, and it is not an impactogen. |
| Windley (1989) | **AMCG Magmatism in Plate-Tectonic Context:** Uses analogies with Phanerozoic settings to develop models for AMCG-related magmatism across Grenville Province. In Ontario, relies largely on 1986 model of Windley. Explains the Midcontinent Rift as a collision-related extension zone analagous to the Rhine Graben. |
| Anderson and Burke (1983) | **Wilson Cycle:** Analysis of Grenville Province geology in Wilson Cycle context. Suggest several sutures may be present in Grenville Province, Midcontinent Rift is impactogen related to continental collision, at least 3 orogenic events may be present in Grenville Province. |
| Farquhar and Fletcher (1980) | **Wilson Cycle:** Lead isotope data consistent with Central Metasedimentary Belt developing in an arc followed by subsequent continental collison. Specific model not developed. |
| Patchett et al. (1978) | **Wilson Cycle Opening and Closing:** Uses paleomagnetic data and Rb-Sr ages on basaltic rocks in North America and Scandinavia to suggest ocean opening and closing to cause Grenville Orogeny. |
| McWilliams and Dunlop (1978) | **Continental-Collision Model:** Uses paleomagnetic data primarily to invoke a model of continental thickening, metamorphism and deformation resulting from continental collision and lateral transport of continental blocks. |
| Irving and McGlynn (1976) | **Wilson Cycle Opening and Closing:** Uses paleomagnetic data to invoke a model of ocean opening and closing to form Grenville Province. Model dependent on paleomagmetic data which was subsequently shown to record mainly late- or post-Grenville magnetizations (*see* Paleomagnetism). Suture of Wilson Cycle would lie south of Central Metasedimentary Belt Boundary Zone in Ontario. |
| Dewey and Burke (1973) | **Continental-Collision Model:** Draws on comparison with Himalayas and Tibet. Explains Grenville Province as a consequence of thickened continental crust and resultant heating and magmatism. Comparison with Himalayas and Tibet still widely used. Model dated only in sense that our knowledge of plate interactions has advanced considerably in the 20 years since the model was proposed. Suture would lie beneath the Appalachians. |
| Donaldson and Irving (1972) | **Rifting and Continental collision:** Proposes rifting of shield followed by continental collison to form orogenic belt. Grenville Front is suture zone. |
| | **Central Gneiss Belt** |
| Dickin and McNutt (1989a, 1990) | **Island-Arc Accretion and Continental Collision:** Nd-Sm isotope studies suggests a Penokean-age suture separates Nipissing and Algonquin terranes. Authors invoke model of northwest- and southeast-directed subduction to form different Central Gneiss Belt terranes and then collison with Central Metasedimentary Belt results in Grenville Orogeny northwest-directed thrusting. Model covers post–1900 to 1100 Ma period of gneiss belt history, Ontario specific. |
| Davidson et al. (1982),<br>Davidson (1984a, 1984b) | **Northwest-directed Thrusting as a Consequence of Continental Collision:** Model of northwestward-directed thrusting as a result of continental collision is invoked to explain terrane distribution in Central Gneiss Belt. Model is regional and conceptual in scope, not specific. Model is concerned with post 1200 million-year-old history of gneiss belt. Most data is in Ontario, although concept is not Ontario specific. |
| | **Central Metasedimentary Belt** |
| Lumbers et al. (1990) | **Mantle-activated Rift Model:** Authors favour mantle-activated rift model of Condie (1982) to explain Central Metasedimentary Belt history. Note similarity to Archean greenstone belt evolution. Authors would place Grenville Orogeny at 1240–1180 Ma (major tectonothermal event), with extension-related plutonism at 1090 Ma and fenite-carbonatite-pegmatite tectonothermal event at 1070–1020 Ma. Explains many aspects of Central Metasedimentary Belt geology. Model is based on Ontario data and is Ontario specific. |

**Table 19.18.** Summary of tectonic models proposed to explain Grenville Province evolution in Ontario.

| | |
|---|---|
| Corriveau (1989, 1990), Corriveau et al. (1990) | **Arc-Continent Collision:** Uses analogy with Sunda Arc to explain origin of late potassic magmatism in Central Metasedimentary Belt and post-Elzevir Terrane, terrane accretion and magmatism. Uses Ontario data and has much relevance to Ontario. |
| Windley (1986) | **Island-Arc Accretion by Analogy with Western Himalaya:** Uses analogies with western Himalayas to propose model of island-arc development followed by collision between Central Gneiss Belt and/or Central Metasedimentary Belt and Central Granulite Terrane. Model is Ontario specific. |
| Gauthier and Brown (1986) | **Aulocogen Model:** Uses setting of zinc deposits and regional geology to suggest a rift (aulocogen) model for Central Metasedimentary Belt. Based mainly on Quebec data but applied to Ontario. Model may have greater relevance to Mont-Laurier Terrane than to Central Metasedimentary Belt terranes in Ontario. |
| Condie and Moore (1977), Bartlett (1983), Smith and Holm (1990a, 1990b), Holm et al. (1986) | **Marginal Basin and/or Back-Arc Models:** Authors use geochemistry on metavolcanic suites to argue for marginal basin and/or back-arc settings for volcanism in western and central Elzevir Terrane. Models use Ontario data and are Ontario specific. Models more conceptual than specific although data is cornerstone of other more elaborate models (e.g., Windley 1986). |
| Brown et al. (1975) | **Island-Arc Model:** Often quoted, this model proposes the existence of relict oceanic crust in the Ontario Central Metasedimentary Belt (Kaladar Complex) and the presence of an adjacent island-arc complex in the present Mazinaw Terrane. Many elements of this model are still invoked today, although some of the data is now disputed (e.g., Kaladar Complex as an ophiolite). Ontario data and is Ontario specific. |
| | **Alternative (Non-Plate-Tectonic) Models** |
| | **Grenville as a Whole** |
| Baer (1981b) | **"Plate-Jam" Model:** Proposes that Proterozoic tectonic regimes were characterized by limited or no subduction. Consequently, orogenesis is related to prolonged stress and heating caused by prolonged continental-collision somewhat, but not directly, analogous to Himalayas and Tibet. |
| Baer (1977) | **"Mega-Shear" Model:** Proposes that there was no subduction in the Proterozoic. Plate boundaries are zones of extensive, prolonged shearing, which are capable of orogenesis. Grenville Province can be viewed as resulting from a "mega-shear" zone. |
| Baer (1976) | **"Aulacogen" Model:** Grenville Supergroup deposited in aulacogen formed between Central Gneiss Belt and Central Granulite Terrane in Quebec with a triple-junction south of Lake Ontario related to development of a proto-Atlantic Ocean. Volcanism is rift-related. Orogeny related to closure of proto-Atlantic with suture lying south of exposed Grenville Province. Relies on Ontario geology although not Ontario specific. |
| Wynne-Edwards (1976) | **"Millipede" Model:** Proterozoic plates are more ductile than in Phanerozoic. Plates drift slowly over lines of hot spots ("hot lines") causing crustal heating, deformation, magmatism and high-grade metamorphism resulting in extensive crustal reworking. |
| Wynne-Edwards (1972) | **Geosynclinal Model:** Although no model is specifically developed, discussion in text alludes to a geosynclinal environment. |

## 1450 TO 1420 MILLION-YEAR-OLD ACTIVITY

Major felsic magmatism of AMCG type (Lumbers and Vertolli 1991) occurred throughout the Central Gneiss Belt in the interval 1450 to 1420 Ma, particularly within the Algonquin Terrane. Associated with this magmatic event was a regional metamorphism, locally reaching granulite-facies conditions (Culshaw et al. 1991). Effects of this magmatic-metamorphic event are recorded in the Grenville Front region near Killarney, where pegmatite dikes and leucosome pods in migmatites of *circa* 1400 million year age associated with regional metamorphism are present (Krogh 1989), and in the Britt Domain (Culshaw et al. 1991). Elsewhere, this earlier granulite metamorphic event is overprinted and largely obliterated throughout much of the Central Gneiss Belt (Culshaw et al. 1991).

Nd-model ages (Dickin and McNutt 1990) suggest that the Parry Sound Terrane consists predominantly of material derived from the mantle at *circa* 1400 Ma. This is consiste[illegible]th the age of the McKellar Diorite of 1425±75 Ma (v[illegible]Breemen et al. 1986). The Parry Sound Terrane may represent the basal portion of an island-arc complex (Easton 1989d; Culshaw et al. 1990). Lithologically, parts of the Grimsthorpe Domain in the Central Metasedimentary Belt resemble the Parry Sound Terrane. At present, however, it appears that the Grimsthorpe Domain is some 100 million years younger than the Parry Sound Terrane, and thus, there is no direct link between the Parry Sound Terrane and the Central Metasedimentary Belt. The only other region where Parry Sound Terrane-type rocks are known occurs in the subsurface in southwestern Ontario in the Cambridge Domain (*see* Figure 19.94).

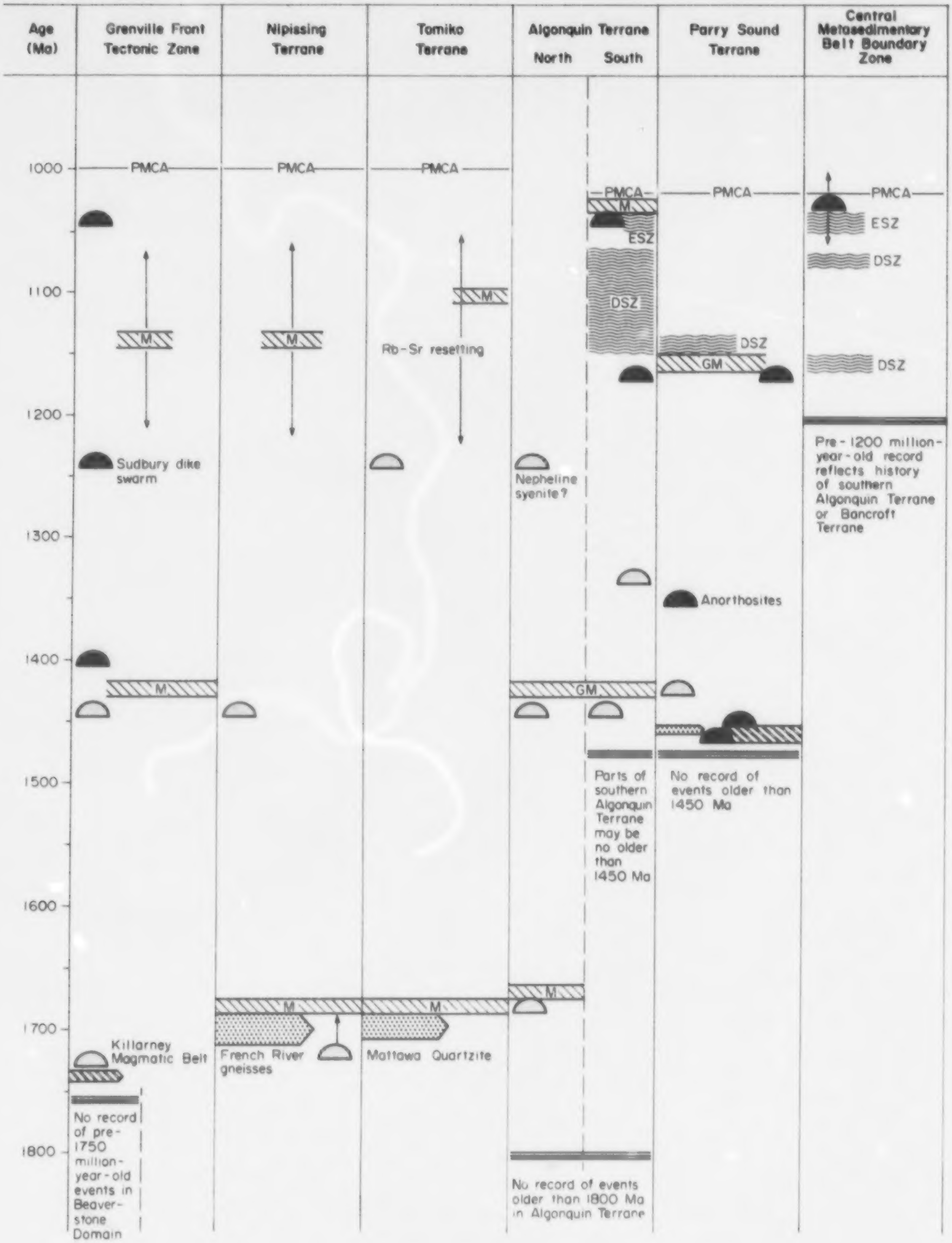

Age (Ma)
Grenville Front Tectonic Zone
Nipissing Terrane
Tomiko Terrane
Algonquin Terrane
North
South
Parry Sound Terrane
Central Metasedimentary Belt Boundary Zone
1000
1100
1200
1300
1400
1500
1600
1700
1800
PMCA
M
ESZ
DSZ
GM
Rb-Sr resetting
Sudbury dike swarm
Nepheline syenite?
Pre-1200 million-year-old record reflects history of southern Algonquin Terrane or Bancroft Terrane
Anorthosites
Parts of southern Algonquin Terrane may be no older than 1450 Ma
No record of events older than 1450 Ma
Killarney Magmatic Belt
French River gneisses
Mattawa Quartzite
No record of pre-1750 million-year-old events in Beaverstone Domain
No record of events older than 1800 Ma in Algonquin Terrane

**Figure 19.95.** Sketch summarizing time-space relationships within the Central Gneiss Belt.

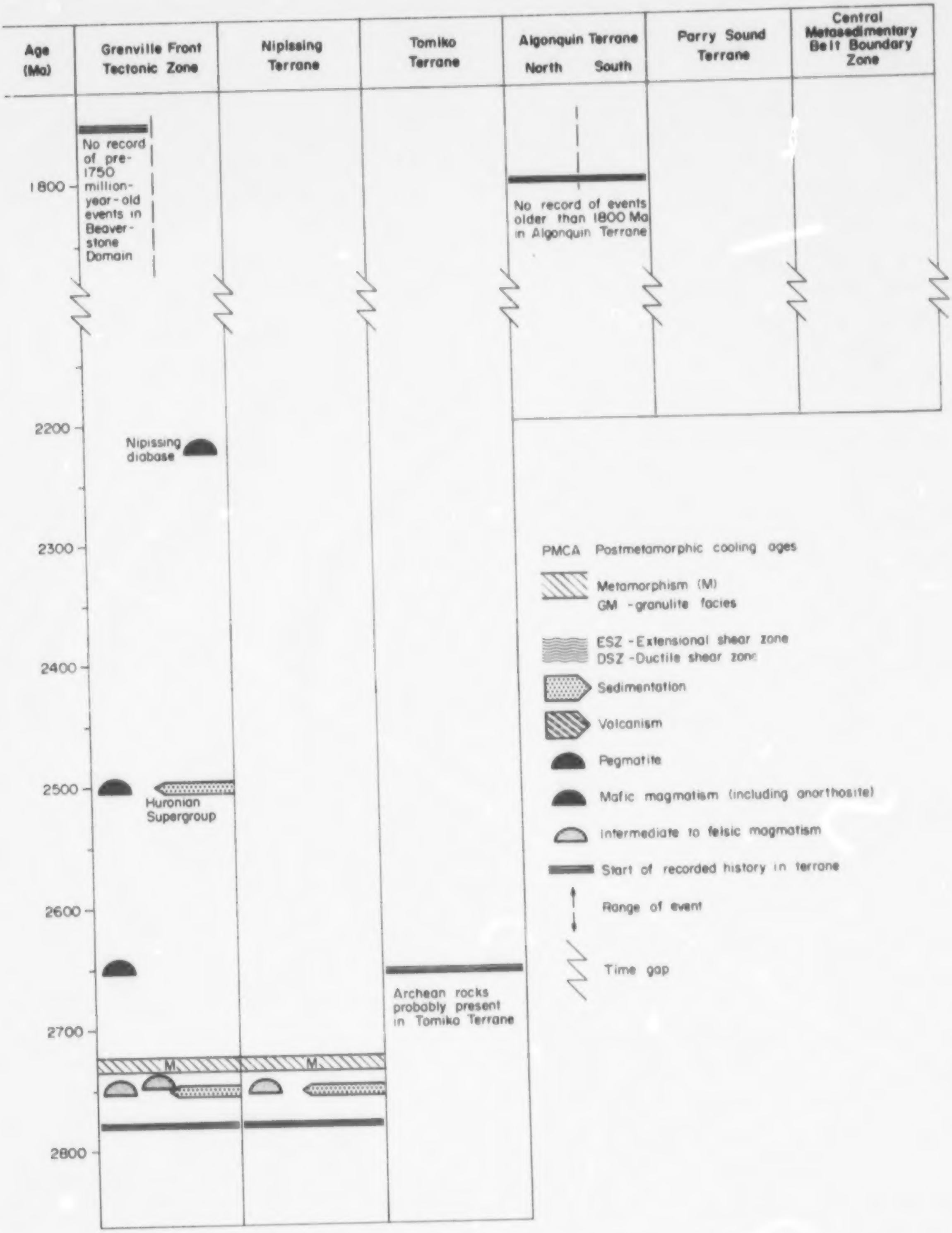


**Figure 19.95.** Sketch summarizing time-space relationships within the Central Gneiss Belt.

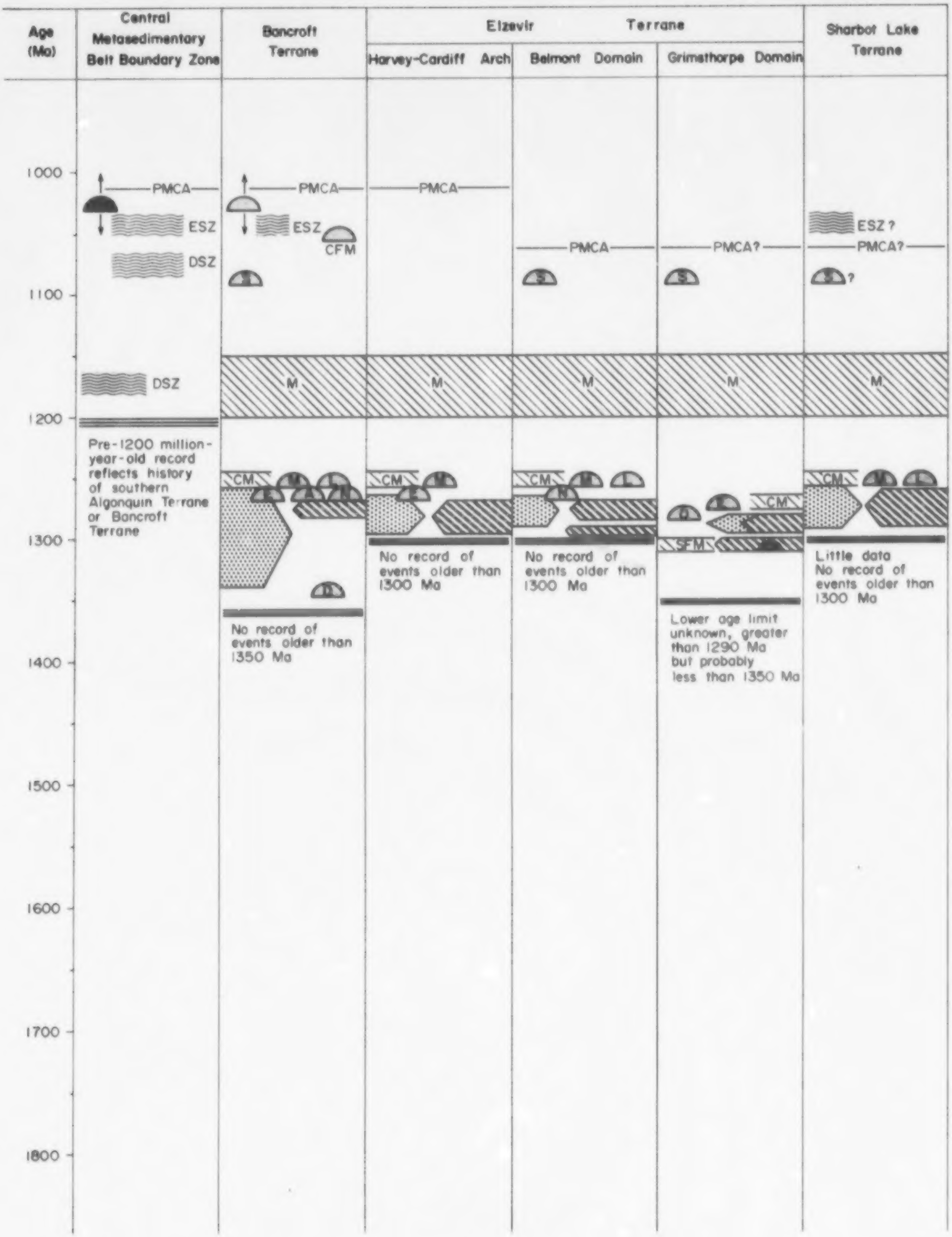


**Figure 19.96.** Sketch summarizing time-space relationships within the Central Metasedimentary Belt.

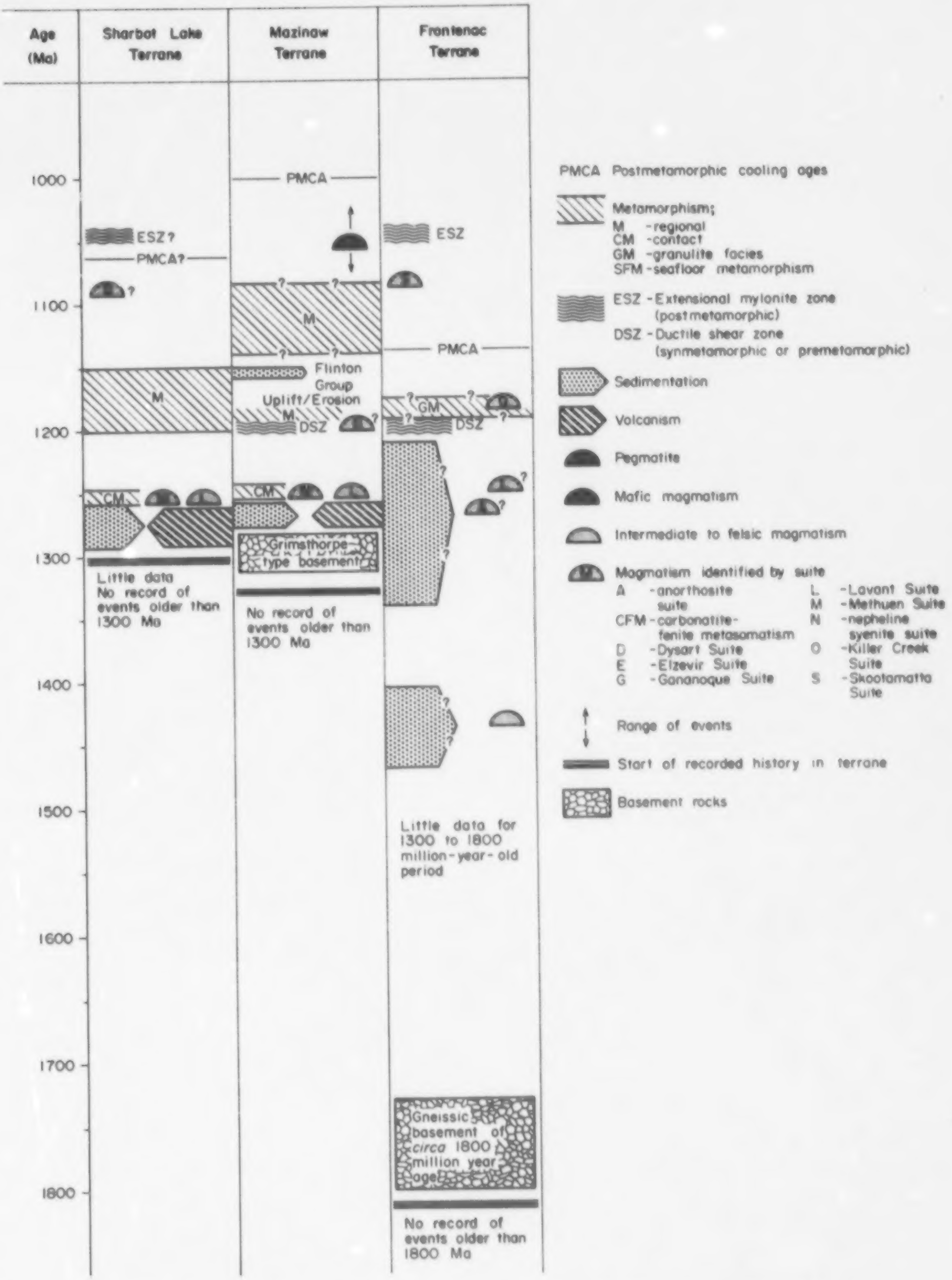

Age (Ma)
Sharbot Lake Terrane
Mazinaw Terrane
Frontenac Terrane
1000
1100
1200
1300
1400
1500
1600
1700
1800
ESZ?
PMCA?
M
CM
Little data
No record of events older than 1300 Ma
PMCA
M
Flinton Group
Uplift/Erosion
DSZ
CM
Grimsthorpe-type basement
No record of events older than 1300 Ma
ESZ
PMCA
GM
DSZ
Little data for 1300 to 1800 million-year-old period
Gneissic basement of circa 1800 million year age
No record of events older than 1800 Ma
PMCA Postmetamorphic cooling ages
Metamorphism;
M -regional
CM -contact
GM -granulite facies
SFM -seafloor metamorphism
ESZ - Extensional mylonite zone (postmetamorphic)
DSZ - Ductile shear zone (synmetamorphic or premetamorphic)
Sedimentation
Volcanism
Pegmatite
Mafic magmatism
Intermediate to felsic magmatism
Magmatism identified by suite
A -anorthosite suite
CFM - carbonatite-fenite metasomatism
D -Dysart Suite
E -Elzevir Suite
G -Gananoque Suite
L -Lavant Suite
M -Methuen Suite
N -nepheline syenite suite
O -Killer Creek Suite
S -Skootamatta Suite
Range of events
Start of recorded history in terrane
Basement rocks

**Figure 19.96.** Sketch summarizing time-space relationships within the Central Metasedimentary Belt.

Evidence of *circa* 1420 million-year-old magmatism (Wellesley Island leucogranite gneiss, McLelland, Chiarenzelli et al. 1988) is found within the Central Metasedimentary Belt in the Frontenac Terrane.

## 1350 TO 1320 MILLION-YEAR-OLD ACTIVITY

A second pulse of magmatism, dominated by tonalitic magmas, occurred in the southern Central Gneiss Belt and farther to the southeast, in the *circa* 1350 to 1320 million year interval. Anorthositic magmas in the Parry Sound Terrane were also emplaced into the crust at this time. Rocks of this age may be basement to parts of the Central Metasedimentary Belt. The age similarity between these plutons and volcanism and sedimentation in the Central Metasedimentary Belt means that basement of this age beneath the Central Metasedimentary Belt would be isotopically transparent.

## 1300 TO 1240 MILLION-YEAR-OLD ACTIVITY

The early development of the Central Metasedimentary Belt, particularly the Bancroft and Elzevir terranes, is particularly difficult to establish, especially the relationship between the different domains within the Elzevir Terrane. The situation is analogous to that observed in Archean greenstone belts of the Superior Province (Williams et al., this volume), where it is now recognized that the greenstone belts are composed of several different volcano-sedimentary assemblages ("domains") in fault contact with one another. Although different tectonic settings can be determined for these assemblages ("domains"), their

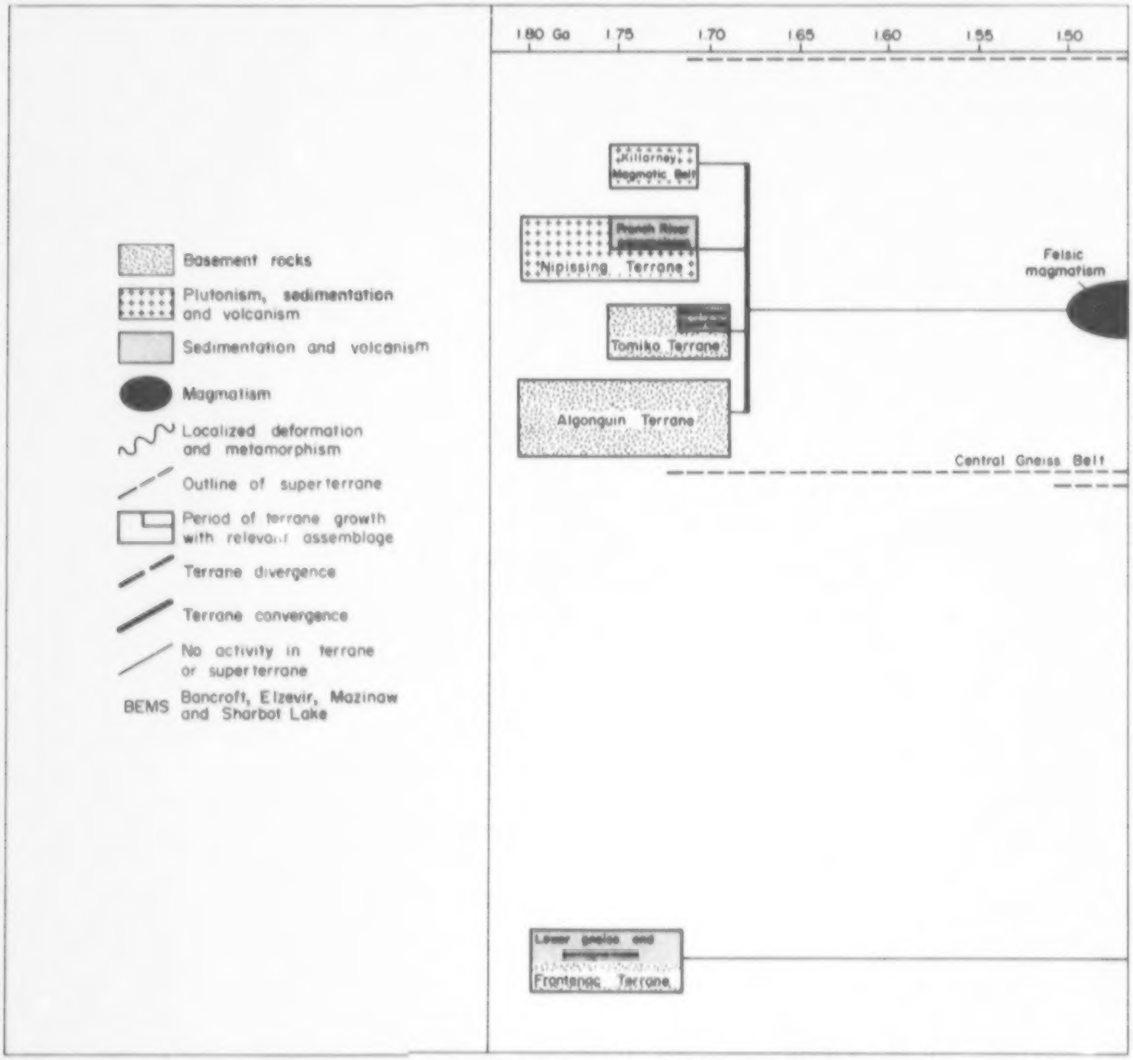

Basement rocks
Plutonism, sedimentation and volcanism
Sedimentation and volcanism
Magmatism
Localized deformation and metamorphism
Outline of super terrane
Period of terrane growth with relevant assemblage
Terrane divergence
Terrane convergence
No activity in terrane or super terrane
BEMS Bancroft, Elzevir, Mazinaw and Sharbot Lake
1.80 Ga
1.75
1.70
1.65
1.60
1.55
1.50
Killarney Magmatic Belt
French River
Nipissing Terrane
Tomiko Terrane
Algonquin Terrane
Felsic magmatism
Central Gneiss Belt
Lower gneiss and
Frontenac Terrane

**Figure 19.97.** Cladogram illustrating the tectonic evolution of the Grenville Province in Ontario.

assembly into belts is still problematic (Williams et al., this volume). Condie and Moore (1977) and Condie (1982) drew attention to the similarity between the Elzevir Terrane and Archean greenstone belts, and if both represent the remnants of complex arc-marginal basin accretion events, then the analogy may be a valid one.

Similarly, within the Elzevir Terrane, a slightly older Grimsthorpe Domain, composed of tholeiitic mafic volcanic and volcaniclastic rocks (ca. 1300 to 1280 Ma) intruded by tonalitic (ca. 1270 Ma) plutonic complexes, may have developed in a marginal basin or back-arc setting. The slightly younger (ca. 1260 to 1250 Ma), more diverse mafic-felsic volcanic and siliceous clastic and carbonate supracrustal assemblage (Belmont Domain, Harvey–Cardiff Arch and the Sharbot Lake Terrane?) is underlain by an older mafic volcanic sequence and could have been constructed on a Grimsthorpe Domain-type basement. The younger supracrustal sequence may represent a marginal basin setting, perhaps formed adjacent to a craton, but more likely adjacent to a "continental fragment." The Bancroft Terrane could represent the platform of this, recently formed, "continental" fragment. The younger supracrustal sequence is not intruded by typical arc plutons, but rather it is intruded by diorite-gabbro (Lavant Suite) and alaskitic granites (Methuen Suite; both ca. 1245 to 1240 million years in age), the latter showing geochemical characteristics typical of anorogenic magmatism (Lumbers et al. 1990). Whether these relationships represent a relatively intact sequence with only minor thrusting that serves to expose different structural levels, or a sequence of accreted terranes which may have been separated from one another by hundreds of kilometres when they developed (perhaps as different sections of a more extensive continental margin) is undetermined (perhaps analogous to the development of the

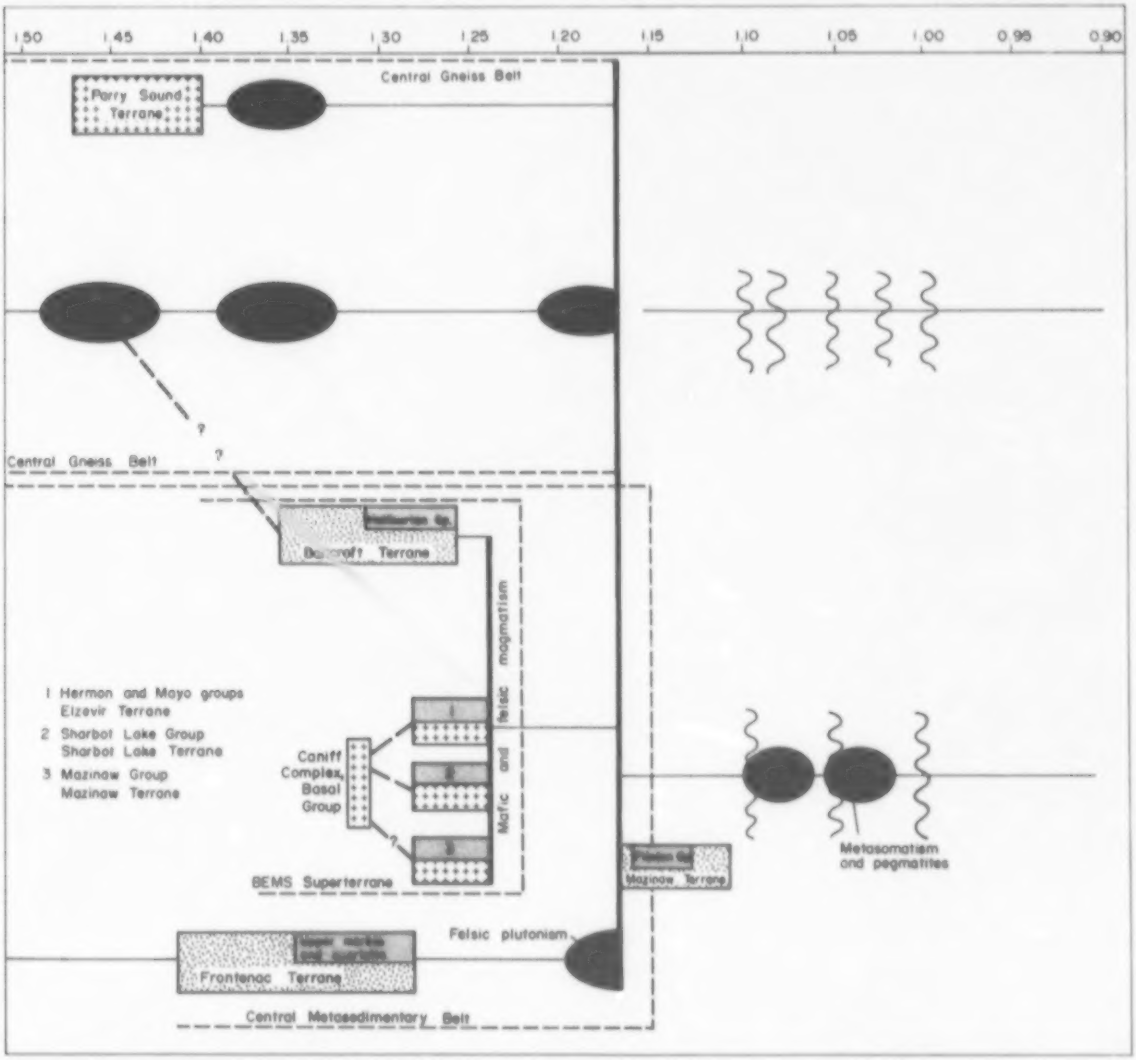

1.50
1.45
1.40
1.35
1.30
1.25
1.20
1.15
1.10
1.05
1.00
0.95
0.90
Central Gneiss Belt
Parry Sound Terrane
?
?
Central Gneiss Belt
Bancroft Terrane
Mafic and felsic magmatism
1 Hermon and Mayo groups
Elzevir Terrane
2 Sharbot Lake Group
Sharbot Lake Terrane
3 Mazinaw Group
Mazinaw Terrane
Caniff Complex, Basal Group
1
2
?
3
BEMS Superterrane
Mazinaw Terrane
Metasomatism and pegmatites
Felsic plutonism
Frontenac Terrane
Central Metasedimentary Belt

**Figure 19.97.** Cladogram illustrating the tectonic evolution of the Grenville Province in Ontario.

Cordilleran margin during the Neoproterozoic to Ordovician). As the Bancroft, Elzevir, and, to some extent, the Sharbot Lake Terrane all contain intrusions of the same plutonic suites (nepheline syenite, anorthosite, Elzevir, Lavant, Methuen), especially the Lavant and Methuen suites, it is likely that the process of amalgamation of the Bancroft and Elzevir terranes into a coherent "superterrane" was largely complete by *circa* 1250 to 1240 Ma, with the magmatism represented by the Lavant and Methuen suites representing the crustal stabilization event. Although these terranes may have been accreted to one another by this time, the resulting "superterrane" may still have been separate from Laurentia.

The geologic setting of the Bancroft and Elzevir terranes in the 1300 to 1240 million-year-old interval may be analogous to a modern setting observed in Indonesia in the vicinity of the Sunda and Banda arcs. Figure 19.98, from McCaffrey and Abers (1991), illustrates the geology of Indonesia 10 million years into the future when orogenesis is complete. Figure 19.98 also illustrates the Central Metasedimentary Belt at a comparable scale. Figure 19.99 illustrates stratigraphic sections from Deep-Sea Drilling Project (DSDP) holes in the region. These sections are roughly comparable to the sections across the Central Metasedimentary Belt shown in Figure 19.54. In Figure 19.98, the final position of the crustal segments is dictated by translation along great distances by strike-slip faults, even though geologic structures will be dominated by the more localized effects of smaller scale thrust faulting. In a complex orogen, the effects of these strike-slip faults may go unrecognized. In the Central Metasedimentary Belt, domain boundaries within the Elzevir Terrane may be far more significant features than presently realized. Figure 19.98 also shows a relatively simple orogenic scenario some 10 million years in the future. Further orogenic activity occurred at several times in the Grenville Province in the interval 1240 to 1070 Ma, consequently the projection in Figure 19.98 some 10 million years into the future, provides only a brief glimpse of the eventual tectonic configuration of the Indonesian region once orogenesis is complete. Figure 19.98 illustrates, simplistically, one scenario for the development of Bancroft- Elzevir Terrane

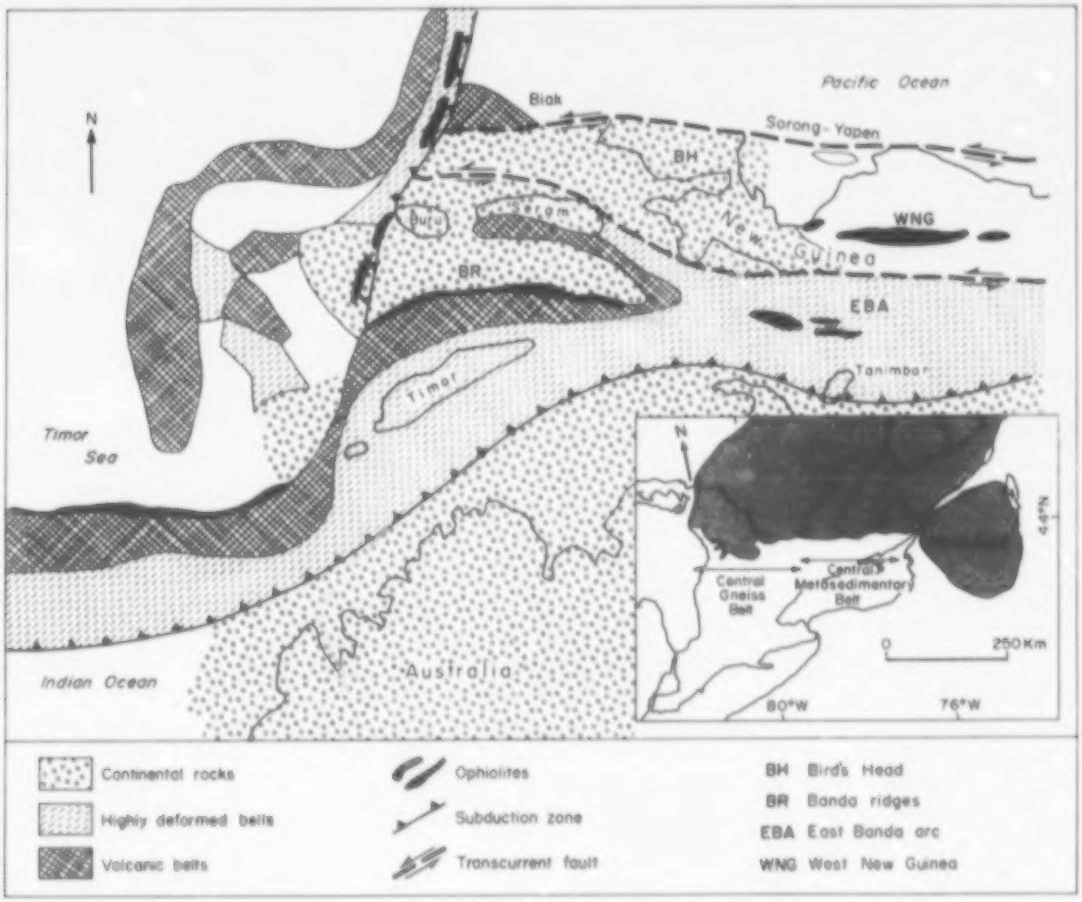

N
Biak
Pacific Ocean
Sorong - Yapen
BH
Seram
Buru
New Guinea
WNG
BR
EBA
Timor
Tanimbar
Timor Sea
Indian Ocean
Australia
N
44°N
Central Gneiss Belt
Central Metasedimentary Belt
0
250 Km
80°W
76°W
Continental rocks
Highly deformed belts
Volcanic belts
Ophiolites
Subduction zone
Transcurrent fault
BH Bird's Head
BR Banda ridges
EBA East Banda arc
WNG West New Guinea

**Figure 19.98.** Hypothetical geologic map of the Banda arc–western New Guinea collision zone 10 my from present (*modified from* McCaffery and Abers 1991). Inset map shows Grenville Province (screened area) in Ontario at same scale.

relationships prior to, and during, amalgamation of the 2 terranes at *circa* 1240 Ma.

Mazinaw Terrane history is more difficult to decipher than the other Central Metasedimentary Belt terranes. The terrane consists of a volcanic arc (ca. 1250 Ma?) and a related siliciclastic and carbonate sediment sequence, that appears to be constructed on Grimsthorpian-type basement rocks (ca. 1270 million-year-old Elzevir Suite intrusions such as the Northbrook and Cross Lake tonalites and the Mellon Lake Complex). It is then intruded by Methuen Suite intrusions such as the Addington Granite (ca. 1240 Ma). At this point, the history of the terrane diverges from that of the Elzevir and Bancroft terranes. The Mazinaw Terrane is subsequently subjected to a high-grade (possibly granulite) metamorphic event, rapidly uplifted and eroded and overlain by the Flinton Group sediments. These in turn are metamorphosed by a regional metamorphic event dominated by high temperatures and slow cooling, with minimal compression, resulting in large, randomly oriented porphyroblasts in most supracrustal units in the Mazinaw Terrane. The early supracrustal and plutonic rocks are roughly analogous to the Elzevir Terrane, but the metamorphic and deformation history and style are similar to the Frontenac Terrane. The Mazinaw Terrane includes a calc-alkalic volcanic sequence, in contrast to the predominantly tholeiitic sequences within the Elzevir Terrane. Many granitoid intrusions in the Mazinaw Terrane are undated; 1180 to 1160 million-year-old plutons of the Gananoque Suite may be present within the Mazinaw Terrane.

Carbonate and clastic sedimentation during the interval 1300 to 1240 Ma may have occurred in the Frontenac Terrane (Upper Marble and Quartzite unit).

The Central Gneiss Belt also locally exhibits evidence of magmatic activity in the 1300 to 1240 million year interval, possibly in an extensional regime. Events include emplacement of the 1238 million-year-old

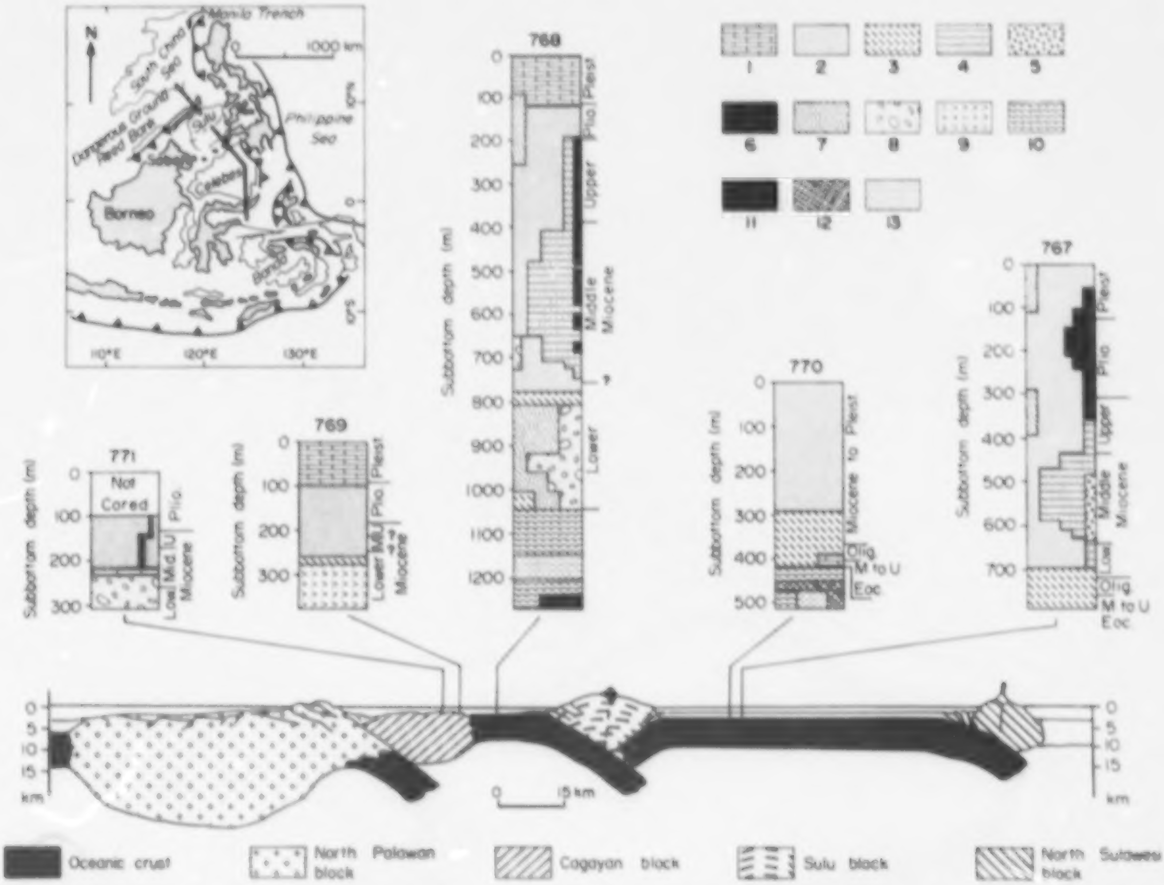

Manila Trench
N
1000 km
South China Sea
Dangerous Grounds
Reed Bank
Sulu
Philippine Sea
Sabah
Celebes
Borneo
Banda
10°N
0°
10°S
110°E
120°E
130°E
768
769
770
771
767
Not Cored
Subbottom depth (m)
Pleist
Plio
Upper
Middle Miocene
Lower
Mid
U
Low
Miocene to Pleist
Olig
M to U Eoc
1
2
3
4
5
6
7
8
9
10
11
12
13
0
5
10
15
km
0
15 km
Oceanic crust
North Palawan block
Cagayan block
Sulu block
North Sulawesi block

**Figure 19.99.** Stratigraphic sections across the Sunda–Banda Arc based on the Deep Sea Drilling Program (DSDP) drill holes (Leg 124 scientific drilling party 1989). Numbers 767 to 771 refer to drill-hole sites. Location of crustal section on inset map. Symbols for stratigraphic columns: 1) nannofossil marl or nannofossil-foraminifer marl; 2) hemipelagic sediments including clay and silt (stone); 3) pelagic brown claystone; 4) terrigenous turbidites; 5) quartz siltstone to sandstone; 6) graded carbonate turbidites; 7) fine ash and tuff; 8) pumiceous, rhyolitic to dacitic coarse tuff and lapillistone; 9) andesitic to basaltic coarse tuff and lapillistone; 10) pillow basalt; 11) basalt sheet flow; 12) brecciated massive basalt; and 13) diabase sill. Compare with Stratigraphic Sections shown in Figure 19.54.

(Krogh et al. 1987) Sudbury dike swarm, emplacement of the alaskitic 1244 million-year-old (Lumbers et al. 1991) Mulock Granite in the Tomiko Terrane and possible nepheline syenite suite magmatism of this age in the Bigwood and Calvin township areas (Nipissing and Algonquin terranes). These first 2 well-dated events correspond in age to Lavant and Methuen suite magmatism, which occurred in a postvolcanic, extensional regime. The nepheline syenite magmatism could be of the same age as that within the Central Metasedimentary Belt and thus would be slightly older (ca. 1290 to 1250 Ma) than the granitic magmatism. The Central Metasedimentary Belt could be separate from the Central Gneiss Belt during this interval (i.e., 1300 to 1250 Ma). Presence of similarly aged magmatic events do not necessarily imply that the Central Metasedimentary Belt and Central Gneiss Belt were contiguous at *circa* 1240 Ma.

## 1180 TO 1160 MILLION-YEAR-OLD ACTIVITY

Not until *circa* 1180 to 1160 Ma are congruent events recorded in the Frontenac Terrane, the rest of the Central Metasedimentary Belt and the Central Gneiss Belt. Thrusting of the Parry Sound Terrane over the Algonquin Terrane and granulite-facies metamorphism in the Parry Sound Terrane occurred at *circa* 1160 Ma (van Breemen et al. 1986). Earliest recorded deformation in the CMBBZ, occurred at *circa* 1180 Ma (McEachern 1990). Within the Central Metasedimentary Belt, Lumbers et al. (1990) suggest that the main period of regional metamorphism related to the "Grenville Orogeny" occurred between 1240 and 1180 Ma, based on ages of plutonic suites in the Central Metasedimentary Belt. This is consistent with observations in the Mazinaw Terrane by Easton and Ford (1991) that a high-grade metamorphism occurred prior to deposition of the Flinton Group, which has a maximum age of 1157±3 Ma (Kinsman and Parrish 1990), and that this earlier metamorphic event overprinted pre-Flinton thrusts in the Mazinaw Terrane. Within the Elzevir Terrane, Berger and York (1981a, 1981b), Lopez-Martinez and York 1983 and York 1984 document a thermal pulse at *circa* 1140 Ma, which interrupts the cooling history of the Elzevir Terrane after emplacement of Lavant Suite diorites and gabbros. Within the Frontenac Terrane, Gananoque Suite magmatism is coincident with granulite-facies metamorphism and occurred between 1175 and 1160 Ma (van Breemen and Davidson 1988b; Lumbers et al. 1990; Marcantonio et al. 1990). All of these ages are consistent with a major orogenic period (referred to as $M_1$ and $D_1$ for sake of discussion below) extending from Parry Sound to Kingston in the interval 1180 to 1160 Ma. This interval would correspond roughly to the termination of the Elzevirian Orogeny (Moore and Thompson 1980; Baer 1981a; Easton 1986a, although not exactly coincident with the ages of the orogeny in these papers). The cause of this regional deformation and metamorphism is not known. In Figure 19.97, it is assumed that it is related to a collisional event between the Central Gneiss Belt and a continental block with a platform consisting of the Frontenac and possibly the Central Granulite terranes. The range in age for this orogenic activity, as well as the scale of crustal thickening and heating of the crust through magmatic activity to generate widespread granulite-facies conditions across the Central Gneiss Belt and the Central Metasedimentary Belt, suggests a continental collisional event, rather than an arc-arc collision.

The Mazinaw and Frontenac terranes may have become accreted to one another during this orogenic episode, as both seem to record similar metamorphic and deformational histories related to the Elzevirian Orogeny. The history of the 2 terranes during the Ottawan Orogeny is not the same, however, as the Mazinaw Terrane is subjected to an amphibolite-facies metamorphic event not preserved in the Elzevir Terrane. This difference might be attributable to thicker crustal stacking over the Mazinaw Terrane during the Ottawan Orogeny, resulting in deeper burial and consequently a higher degree of metamorphism.

In the Grenville Province in Quebec, most anorthosite massifs were emplaced at roughly 1160 million years, probably in an extensional regime (Gower et al. 1990; Higgins and Hervet 1992), indicating that by this time, collision associated with the Elzevirian Orogeny was complete. Within the Central Metasedimentary Belt, extensional events at this time are represented by emplacement of gold-bearing quartz veins in shear zones (*see* Easton and Fyon, this volume) and deposition of the Flinton Group.

Knowledge of events within the Central Metasedimentary Belt between *circa* 1160 and 1090 Ma is limited. In the Mazinaw Terrane, rapid uplift and erosion occurred, followed by Flinton Group deposition in a probable extensional environment. Little other activity is known from either the Central Gneiss Belt or the Central Metasedimentary Belt. Perhaps it may reflect a period of uplift and stabilization.

## 1100 TO 1060 MILLION-YEAR-OLD ACTIVITY

In the interval between *circa* 1100 and 1000 Ma, there may be 2 discrete tectonic episodes which closely follow one another. In the interval between 1100 to 1060 Ma, thrusting within terrane and domain boundaries in the CMBBZ and the Central Gneiss Belt is recorded by syntectonic pegmatites ($D_2$)(e.g., van Breemen and Hanmer 1986; McEachern et al. 1990; Nadeau and van Breemen 1990). This period of thrusting corresponds roughly to the "Ottawan" or the classic "Grenville" orogeny. Granulite-facies metamorphism occurred in the central Algonquin Terrane (van Breemen et al. 1986; van Breemen and Davidson 1990), and retrograde (upper amphibolite metamorphism) is recorded in terrane and domain boundaries. Within the Mazinaw Terrane, upper greenschist- to lower amphibolite-facies metamorphic conditions existed, as the Flinton Group was metamorphosed sometime in the interval *circa* 1140 to 1000 Ma. The regional metamorphism ($M_2$) in the Mazinaw Terrane has no known associated magmatism and thus was probably induced by burial, probably by crustal stacking during the same thrusting event that affected the Central Gneiss Belt. The cause of this event is unknown. It could be due to continent-continent collision that occurred south of the exposed Grenville Province.

Other than in the Algonquin and Mazinaw terranes, there is no indication of a significant regional metamorphic event elsewhere in the Grenville Province during this period, although a low grade thermal event in this interval may have reset rubidium-strontium and potassium-argon isotopic systems across the Grenville Province. In the Belmont Domain, this is recorded by ubiquitous 1070 million-year-old titanite ages (Davis and Bartlett 1988; Lumbers et al. 1990).

Within the Central Metasedimentary Belt, Skootamatta Suite postmetamorphic intrusions were emplaced into the Bancroft, Elzevir and Frontenac terranes between *circa* 1090 to 1075 Ma (Corriveau 1990; Corriveau et al. 1990). Plutons of this suite appear to be absent from the Mazinaw and Sharbot Lake terranes. Postmetamorphic Skootamatta Suite intrusions in the same area have ages of *circa* 1088 Ma, implying that the titanite ages record cooling rather than another thermal pulse. Corriveau et al. (1990) suggest that the ultrapotassic to shoshonitic Skootamatta Suite plutons were emplaced above an active subduction zone that dipped to the northwest beneath the Elzevir Terrane. Magmatism related to this arc, however, consists only of a limited number of plutons, and no other major magmatic event is known from either the Central Metasedimentary Belt or the Central Gneiss Belt in this interval. Keweenawan Rift magmatism did however occur in the interval 1100 to 1085 Ma in the foreland region of the Grenville Orogen at this time, although it cannot be directly related to the ultrapotassic magmatism in the Central Metasedimentary Belt. Anderson and Burke (1983) and Gordon and Hempton (1986) have postulated that the Keweenawan Rift represents an "impactogen" that formed as a result of continental collision, analogous to the Red Sea developing in front of the Zagros Front in the Middle East (Figure 19.100). Corriveau et al. (1990) also suggest that the Elzevir Terrane (defined as the area of the Central Metasedimentary Belt including Skootamatta Suite plutons, and thus different in extent than the Elzevir Terrane shown in Figure 19.51a) was not accreted to the Central Metasedimentary Belt until after *circa* 1075 Ma. Shoshonitic to ultrapotassic magmatism is known from recent posttectonic extensional environments, and in these instances, the geochemical characteristics of these intrusions were inherited from mantle subjected to an earlier subduction-related event (ca. 1270 to 1245 Ma; Corriveau et al. 1990). Corriveau et al. (1990) suggest mantle inheritance is unlikely given the extreme geochemical features of this plutons.

During this interval, the architecture of the Central Metasedimentary and Central Gneiss belts observed today was probably developed. Ductile thrusting resulted in stacking and disruption of early formed "superterranes." The Elzevir, Bancroft and Sharbot Lake terranes were thrust over the Frontenac and Mazinaw terranes. Subsequent extension along fault boundaries resulted in considerable vertical movement and local rapid erosion, which resulted in the terrane and domain boundaries currently observed in the Grenville Province. For example, the Frontenac Terrane was probably unroofed and uplifted rapidly, whereas the Mazinaw Terrane was buried for longer and uplifted slowly, thereby suffering more intense metamorphism during the Ottawan Orogeny.

During this interval, a period of thrusting and an arc-related magmatic event are recognized. The cause and the configuration of the various terranes of the Central Gneiss Belt and Central Metasedimentary Belt are undetermined.

## 1060 TO 1000 MILLION-YEAR-OLD ACTIVITY

Following the period of compressional activity between 1100 to 1060 Ma, extensional events occurred throughout the Central Metasedimentary Belt for roughly 60 million years, particularly in the Bancroft Terrane, where they were manifested by carbonatite-fenite suite magmatism and metasomatism and granite pegmatite injection. During this interval, the area was also subjected to cooling, which occurred at different rates across the Central Metasedimentary Belt and the Central Gneiss Belt, probably related to extensional gravitational collapse of the thrust stack assembled during the Elzevirian and Ottawan orogenies (cf., Cosca 1989; Carlson et al. 1990; Mezger et al. 1991; Figure 19.17). By *circa* 1000 Ma, cooling below argon blocking temperatures of hornblende across most of the Central Metasedimentary Belt and Central Gneiss Belt was complete, with cooling through biotite-blocking temperatures completed by 950 Ma. The timing of the "Grenville Event" or "Grenville Orogeny" has long been associated with this late, cooling event, which in essence reflects the end of regional orogenic activity in the Grenville Province; not the peak of orogenesis.

Extensional faulting along domain and terrane boundaries is recorded also within the Central Gneiss Belt in this period (ca. 1040 Ma, Nadeau and van Breemen 1990), and major shear systems such as the RLMZ and the Bancroft shear zone were formed at this time. The Sharbot Lake Terrane was probably emplaced into its present position then and represents a thin crustal sheet (cf., Real and Thomas 1987), probably a sliver of Elzevir Terrane, that was back-thrust and tectonically wedged between the Mazinaw and Frontenac terranes.

## POST-1000 MILLION-YEAR-OLD ACTIVITY

Extension continued in the period 1000 to 900 Ma during cooling of the Grenville Orogen, and at *circa* 900 Ma was characterized by mafic dike injection (Kingston Swarm), minor carbonatitic and alkalic magmatism and faulting related to development of a precursor to the Ottawa graben.

From 1000 to 500 Ma, the Grenville Orogen was uplifted and eroded. Much of the detritus from the orogen may have been transported northward into Arctic region sedimentary basins, such as the Late Proterozoic Shaler Basin (e.g., Rainbird et al. 1991). By the Early Paleozoic, the Grenville Orogen was essentially a peneplain.

Extension was renewed in the Late Proterozoic to Early Cambrian, with faulting along the Ottawa–Bonnechere Graben, especially in the Pembroke to Sudbury region, coincident with emplacement of carbonatitic and alkalic complexes in the Nipissing area and Grenville Swarm diabase dikes. This extensional event is probably related to rifting of North America in conjunction with the development of the Iapetus Ocean. During this time, and until the deposition of fluvial and nearshore marine sediments in the Middle Cambrian, the area was in a tropical environment; regoliths and karsts were developed on the exposed Grenville surface.

## DISCUSSION

The history outlined above and shown schematically in Figure 19.97, is a tentative scenario. Figure 19.98 and 19.99 illustrate 2 main points. First, the tectonic evolution of the Central Metasedimentary Belt can be compared with modern arc systems, such as the Sunda and Banda arcs. Previous plate-tectonic models applied to the Grenville Province have taken a more simplistic approach, and although they have served to identify the general tectonic regime of part of the Central Metasedimentary Belt, they have not fully conveyed the complicated histories that exist between the various terranes. Second, the comparison with the Sunda and Banda arcs reveals that the tectonic evolution of such an arc is extremely complicated, and may never be fully unravelled because of the inability to track ancient plate motions.

The analysis above identifies major periods of orogenic activity throughout parts of the Grenville Province. The main difficulty in developing a tectonic synthesis lies in attempting to relate these events to the spatial distribution of the various terrane and domains at various periods in time, especially as there is geologic, seismic and gravity data indicating that the terranes and domains are all allochthonous south of the French River.

For example, does the presence of Skootamatta Suite plutons indicate that the Elzevir Terrane was emplaced in its current position after *circa* 1076 Ma or do they in fact indicate that the Bancroft, Elzevir and Frontenac terranes coalesced prior to *circa* 1090 Ma, and that the plutons were intruded into these terranes after the allochthon was emplaced? Interpretation of the past spatial relationships between the terranes is further complicated by *circa* 1040 million-year-old extensional activity which has modified the cooling history of the Central Metasedimentary Belt, resulting in rapid cooling and uplift in some areas and slow cooling in others (cf., Cosca 1989).

As discussed above, if an Indonesia- or a Cordilleran-type model is considered as an analog for Grenville Province tectonic development, then the various terranes and domains could have undergone considerable translation with respect to each other, yet possess similar stratigraphic and chronologic histories.

This discussion has focussed on the Grenville Province in Ontario. It should be noted however, that the allochthonous nature of the terranes makes correlation of Ontario geology into adjacent areas of Quebec and New York State difficult. Figure 19.101 illustrates potentially equivalent terranes adjacent to Ontario. In the north, the Tomiko Terrane extends several tens of kilometres into a region of Quebec that has not been mapped in detail. Along the Val d'Or–Montréal Highway corridor, some 150 km east, the geology of the Grenville Province (Indares and Martingole 1989, 1990a, 1990b) bears little resemblance to that seen in Ontario, either in terms of rock type, or structural and metamorphic history. Farther south, in the Central Metasedimentary Belt, it is difficult to trace geology across the Ottawa River into the Mont-Laurier region of the Central Metasedimentary Belt (*see* Figure 19.101). The Elzevir, Mazinaw and Sharbot Lake terranes have no clear counterpart in Quebec, and the Bancroft and Frontenac terranes may be juxtaposed in Quebec. Movement along Ottawa graben faults may have enhanced the geologic contrast by exposing different crustal levels on either side of the river. Thus, although development of the Mont-Laurier region is related to overall development of the Central Metasedimentary Belt, the exact relationships between the geology of western Quebec and the Central Metasedimentary Belt in Ontario are unknown at present. Thus, mineral deposit models developed for the Elzevir Terrane may have little application in the Mont-Laurier Terrane, and *vice versa*.

In the description of the Frontenac Terrane, the similarities between the Adirondack Lowlands and the Frontenac Terrane have already been noted. The lowlands, in part, seem to represent a lower metamorphic grade equivalent of the Frontenac Terrane. Adirondack Lowland-Highland relations have been described by many authors (cf., deLorraine and Carl 1986, McLelland, Chiarenzelli et al. 1988; McLelland et al. 1992) and are not elaborated upon

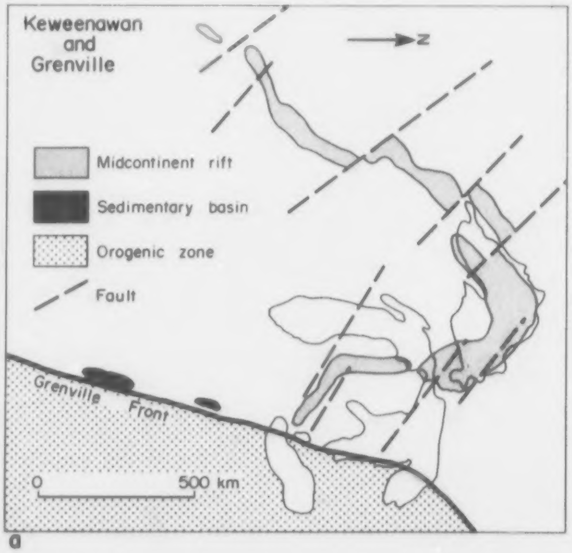


**Figure 19.100.** Tectonic similarity between **(a)** eastern North America at 1100 Ma and **(b)** the modern Middle East. *Modified from* Gordon and Hempton (1986).

herein, other than to say that this plutonic-dominated terrane records magmatism roughly coincident with that seen in Ontario, namely: 1) a *circa* 1300 million-year-old tonalite suite; 2) a *circa* 1250 million-year-old granitic suite; 3) a *circa* 1160 to 1145 million-year-old monzonite-syenite-granite suite; 4) a *circa* 1130 million-year-old granite suite; 5) a 1100 to 1090 million-year-old granite suite; and 6) a 1080 to 1060 million-year-old suite of leucogranite and pyroxene syenites (Chiarenzelli and McLelland 1991). The 1160 to 1145 million-year-old suite is similar to the age of Frontenac Terrane magmatism and is widespread in the highlands and in the Central Granulite Terrane in Quebec (Emslie and Hunt 1990; Higgins and Hervet 1992). Plutons of suite (4) and (5) have not been reported from Ontario, suite (6) corresponds, at least in part, to the Skootamatta Suite. The Adirondack Highlands history seems to be most closely tied to Frontenac Terrane events, although the grouping of plutonic events is roughly coincident with orogenic and magmatic events that are documented across the Grenville Province in Ontario.

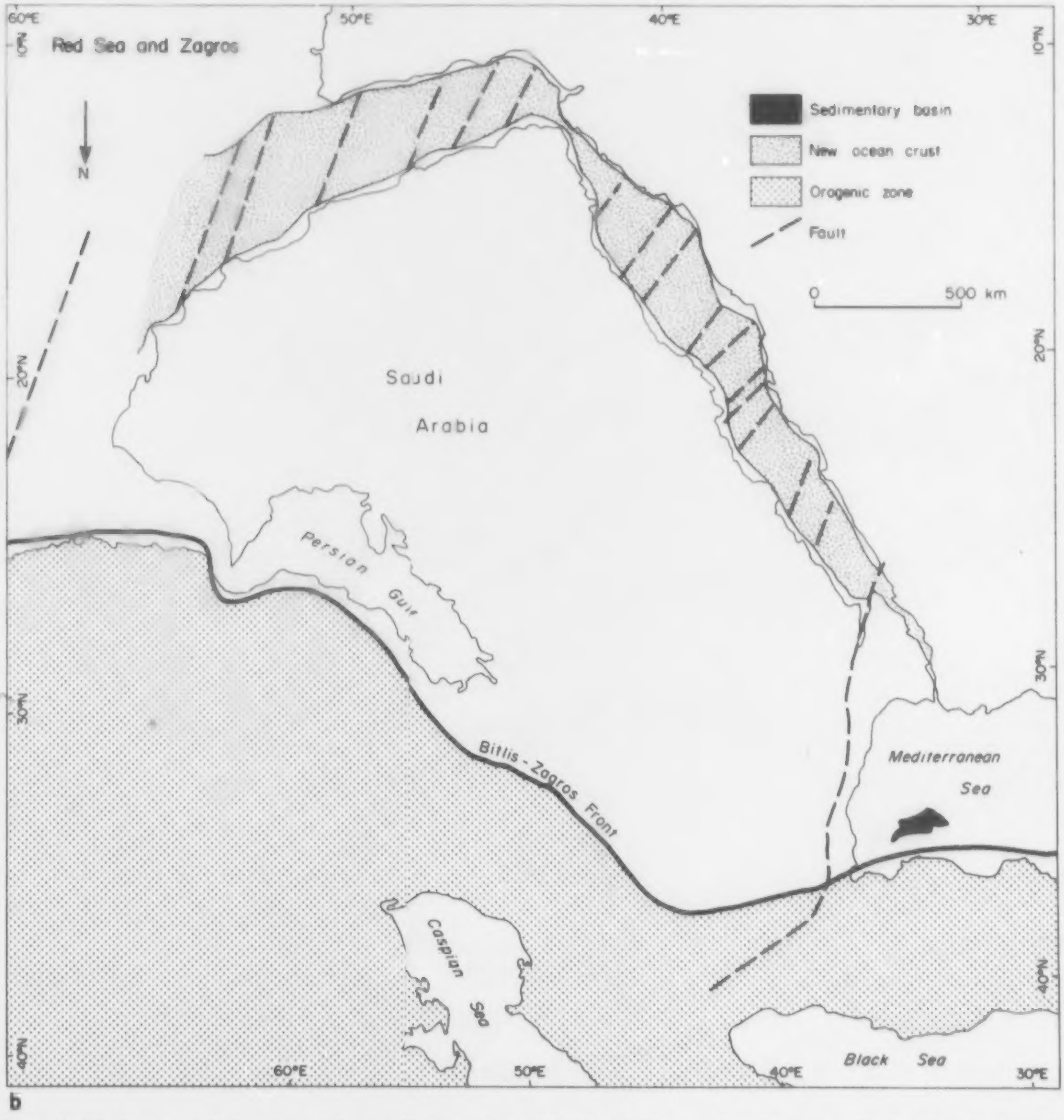

60°E
50°E
40°E
30°E
10°N
Red Sea and Zagros
N
Sedimentary basin
New ocean crust
Orogenic zone
Fault
0
500 km
20°N
Saudi
Arabia
Persian
Gulf
30°N
Bitlis - Zagros Front
Mediterranean
Sea
Caspian
Sea
40°N
Black Sea
b

**Figure 19.100.** Tectonic similarity between (a) eastern North America at 1100 Ma and (b) the modern Middle East. *Modified from* Gordon and Hempton (1986).

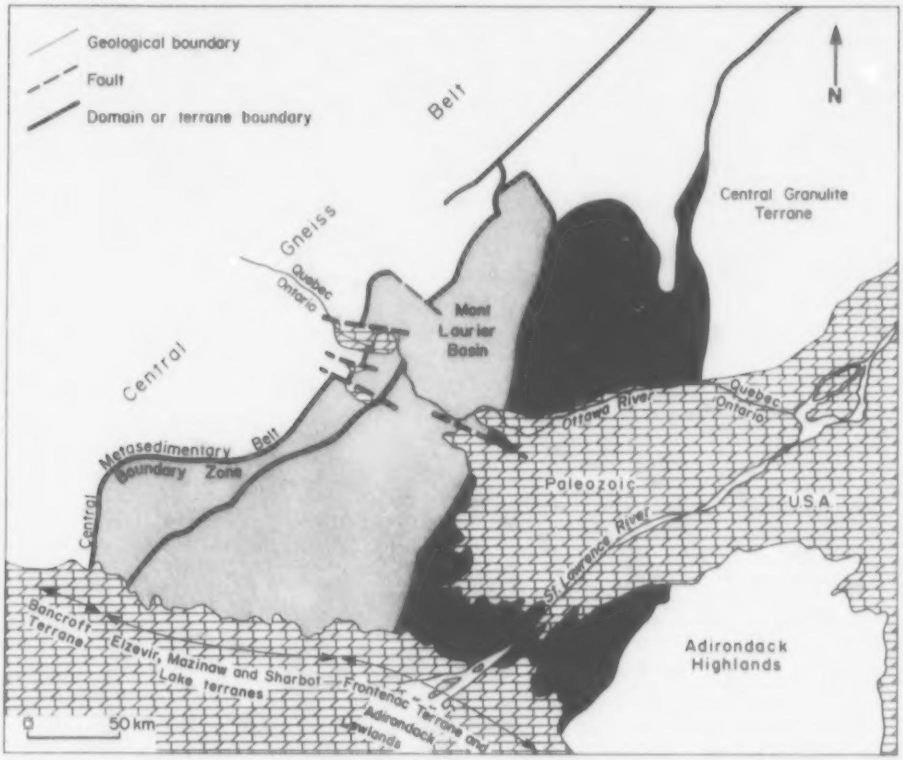

Geological boundary
Fault
Domain or terrane boundary
Belt
N
Central Granulite Terrane
Gneiss
Quebec
Ontario
Mont Laurier Basin
Central
Metasedimentary Belt
Boundary Zone
Central
Ottawa River
Quebec
Ontario
Paleozoic
U.S.A.
St. Lawrence River
Bancroft Terrane
Elzevir, Mazinaw and Sharbot Lake terranes
Frontenac Terrane and Adirondack Lowlands
Adirondack Highlands
0
50 km

**Figure 19.101.** Extension of Central Metasedimentary Belt terranes into western Quebec.

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# Paleozoic and Mesozoic

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# Chapter 20

# Paleozoic and Mesozoic Geology of Ontario

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## Abstract

**The stratigraphy, paleontology and resources of the Paleozoic and Mesozoic bedrock of Ontario have been systematically studied for over a century, especially where these strata are well exposed at surface or encountered in shallow drilling and excavations. More recent investigations have added much to our knowledge of these strata in northern Ontario and in the deep subsurface. This chapter reviews the Paleozoic and Mesozoic stratigraphy and history of Ontario and examines the relationship between sedimentation and tectonics as recorded in this succession.**

**The Paleozoic to Mesozoic succession in Ontario records the periodic inundation of the North American craton by basin-centred inland seas (4 such sedimentary basins are centred in or adjacent to the province). This succession has been subdivided into 10 depositional sequences; defined as stratigraphic intervals of genetically related sedimentary rocks which are commonly bounded by unconformities. The boundaries of some of these sequences correspond to continent-wide sequence boundaries (e.g., Sloss 1963), whereas others are smaller scale and indicate more localized controls.**

**Tectonic controls on sedimentation are evidenced by depositional and erosional patterns. Wedges of orogen-derived clastic rocks correlate with major (and minor) episodes of orogenic uplift. Craton-derived clastic rocks reflect uplift of interbasinal arch areas. Evaporite-dominated intervals indicate basin restriction. Carbonate-dominated intervals indicate relative tectonic quiescence. Erosional patterns vary from craton-wide to arch-centred unconformities, and reflect different scales and styles of cratonic uplift. Local depositional and/or erosional patterns suggest that some cratonic tectonism was accommodated by brittle deformation (e.g., fracture-framework).**

**Correlation of these 10 sequences among the 4 sedimentary basins, in or adjacent to Ontario, affords a comparison of sedimentation in various tectonic settings (e.g., foreland and intracratonic basins, interbasinal arches) and may help to refine constraints on basin development and subsidence models. Some of these models suggest linkage of cratonic and plate margin tectonism. Higher resolution chronostratigraphy or biostratigraphy is required for more detailed correlation of depositional and erosional events on the craton with tectonic events at the plate margin. In addition, more basic data is required from certain stratigraphic intervals and geographic areas (e.g., Hudson Bay Basin) to facilitate more accurate correlations.**

## INTRODUCTION

This chapter provides an overview of Ontario's Paleozoic and Mesozoic geological history. During these eras, large parts of the North American continent were periodically covered by a series of inland seas which left a legacy of carbonate, siliciclastic and evaporite sedimentary deposits. The resultant sedimentary rock record is far from complete, as it contains erosional or nondepositional gaps representing time intervals when the seas had receded. The geological history of this time must then be reconstructed from the remnant rock record.

In spite of periods of erosion, Paleozoic and Mesozoic strata cover approximately 320 000 km² of Ontario, or roughly one-third of the surface area of the province (Figure 20.1), mainly restricted geographically to 2 physiographic areas, the Hudson Bay Lowland of northern Ontario and the St. Lawrence Lowland of southern Ontario. These rocks are the source of a number of valuable commodities in the province, including: oil and natural gas, salt, gypsum, aggregate, cement, building stone, shale clay, chemical grade stone and groundwater. As these strata form the bedrock on which 90% of the province's population lives, interest in these rocks is increasing in light of current environmental and land-use pressures.

### Tectonics and Sedimentation

Regional tectonic events had 2 major influences on the deposition of sedimentary rocks in Ontario during the Paleozoic and Mesozoic eras. Firstly, orogenic activity centred along the present day Appalachian mountain range episodically supplied large amounts of clastic (i.e., siliciclastic or terrigenous) detritus to the area, directly affecting the lithologic character of the deposits. Secondly, tectonic forces promoted and affected the development and positioning on the craton of basins and arches (Figure 20.2; Figure 1, Map 2582, map case). These tectonic elements in turn influenced relative water depth, current flow, salinity, sediment distribution patterns and a variety of other environmental factors in the inland seas.

Two basic types of sedimentary basins formed in or immediately adjacent to Ontario during the Paleozoic (*see* Figure 20.2). The first type, the foreland basin, is represented by only 1 basin in this area, the Appalachian Basin.

This ovoid basin formed cratonward of the various Paleozoic orogenies which occurred along the eastern margin of North America. The basin developed due to lithospheric flexure caused by loading of the crust by emplacement of thrust sheets during orogenic activity (Quinlan and Beaumont 1984; Bally 1989). The remaining sedimentary basins in and around Ontario (*see* Figure 20.2) are considered intracratonic basins, that is, basins founded on the craton and formed by processes not directly related to orogenic activity. The origin of intracratonic basins is less well understood and a number of models have been proposed to explain their formation (Quinlan 1987; *see* Williams, Stott, Thurston et al., this volume).

Subsidence rates for sedimentary basins have varied through time, and also have varied relative to the distance from the basin centre. For example, in the central part of the Michigan Basin during the Middle Ordovician to Early

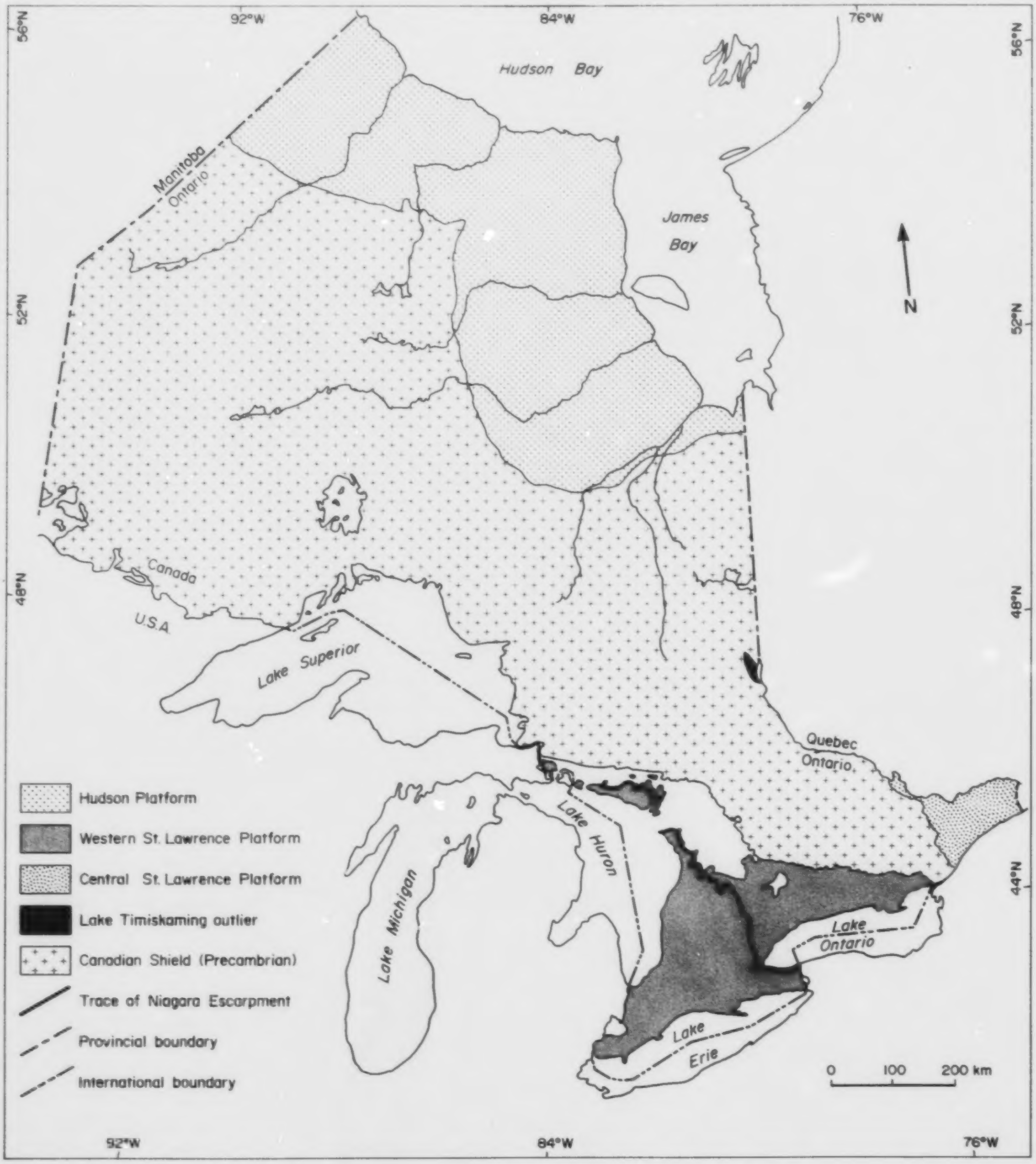
92°W
84°W
76°W
56°N
52°N
48°N
44°N
Hudson Bay
James Bay
N
Manitoba
Ontario
Canada
U.S.A.
Lake Superior
Quebec
Ontario
Lake Huron
Lake Michigan
Lake Ontario
Lake Erie
0
100
200 km
Hudson Platform
Western St. Lawrence Platform
Central St. Lawrence Platform
Lake Timiskaming outlier
Canadian Shield (Precambrian)
Trace of Niagara Escarpment
Provincial boundary
International boundary

**Figure 20.1.** Paleozoic and Mesozoic bedrock platforms in Ontario.

Devonian, the average subsidence rate was 20 to 30 metres per million years, while farther from the basin centre, in southwestern Ontario, the rate was only 5 to 10 metres per million years (Sloss 1988). As the Ontario landmass does not directly overlie a major basin depocentre (except for the relatively small Moose River Basin), the maximum recorded thickness of Paleozoic and Mesozoic sedimentary deposits within the province is only about 1525 m (Winder and Sanford 1972).

The higher subsidence rates of basin depocentres relative to the interbasinal arches resulted in the former accumulating greater thicknesses of sediments than the latter. During periods of cratonic uplift (or epeirogeny) or

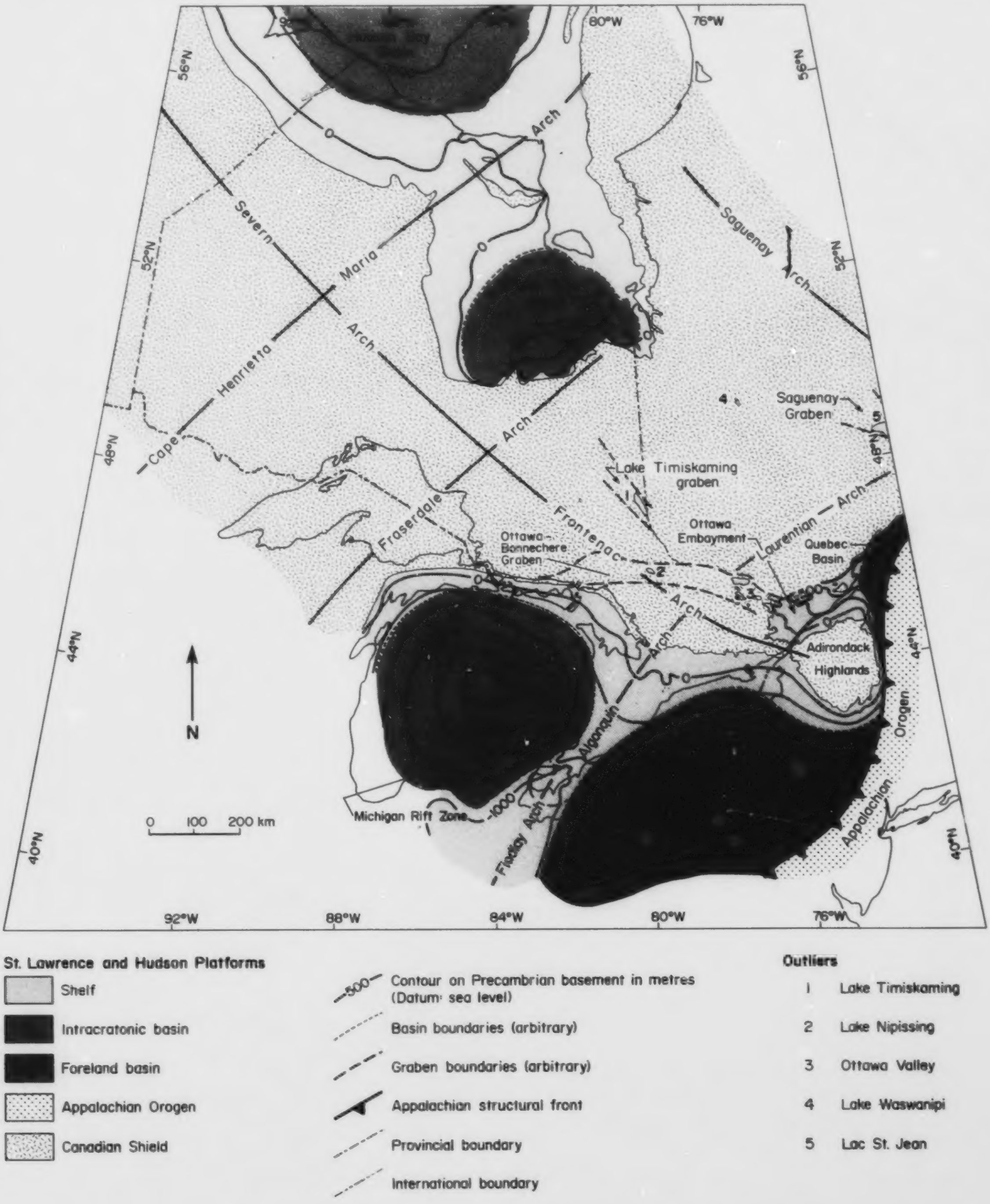
80°W
76°W
56°N
52°N
48°N
44°N
40°N
92°W
88°W
84°W
Arch
Severn
Arch
Maria
Henrietta
Cape
Arch
Fraserdale
Saguenay
Arch
Saguenay Graben
4
5
Lake Timiskaming graben
Ottawa Embayment
Laurentian
Arch
Quebec Basin
Ottawa-Bonnechere Graben
Frontenac
Arch
2
3
1
-500
Adirondack Highlands
Orogen
Appalachian
0
Algonquin
Arch
Findlay Arch
-1000
Michigan Rift Zone
N
0 100 200 km
St. Lawrence and Hudson Platforms
Shelf
Intracratonic basin
Foreland basin
Appalachian Orogen
Canadian Shield
-500
Contour on Precambrian basement in metres (Datum: sea level)
Basin boundaries (arbitrary)
Graben boundaries (arbitrary)
Appalachian structural front
Provincial boundary
International boundary
Outliers
1 Lake Timiskaming
2 Lake Nipissing
3 Ottawa Valley
4 Lake Waswanipi
5 Lac St. Jean

**Figure 20.2.** Principal Paleozoic and Mesozoic tectonic elements of Ontario and adjacent regions.

sea level drop, erosion of sediments would generally be greatest on topographic highs, such as interbasinal arches. These processes have resulted in a less complete sedimentary record being preserved at the basin margins, where these strata typically outcrop today, than in the basin centres.

## Sequence Stratigraphy

Periodically, the inland (or epicontinental) seas receded entirely from the North American continent resulting in widespread subaerial exposure, erosion and the development of continent-wide unconformities (similar to the situation of today). Some workers have attributed the transgression and regression of these epicontinental seas to eustatic, that is, worldwide, fluctuations in sea level. Possible controls on eustatic changes in sea level range from variations in the intensity of worldwide tectonism (i.e., increased seafloor spreading rates) to variations of global climate causing "hot house" or "ice house" (i.e., glaciation) conditions. Widespread regression of epicontinental seas may also be attributed to epeirogenic uplift of the craton.

Sloss (1963) recognized that inter-regional unconformities separated and bounded successions of sedimentary rocks, which he termed "sequences". Sloss subdivided the North American Phanerozoic succession into 6 sequences, and subsequently (Sloss 1988) subdivided these into subsequences. Local or regional (i.e., single basin scale) conditions control relative sea level fluctuations and may result in the superposition of smaller-scale depositional sequences and unconformities on Sloss's larger first or second order sequences. Mitchum et al. (1977, p.53) defined the "depositional sequence" as a "relatively conformable succession of genetically related strata bounded by unconformities and their correlative conformities" and stated (Mitchum et al. 1977, p.55) that depositional sequences were an order of magnitude smaller than Sloss's sequences.

The Paleozoic and Mesozoic stratigraphy of Ontario is herein divided into 10 sedimentary sequences, numbered 1 through 10, and shown in ascending stratigraphic order on Chart D—Paleozoic and Mesozoic Depositional Sequences and Events in Ontario (Map 2582, map case). As illustrated in Figure 20.3, these 10 sequences are generally on a smaller scale compared to the continental-scale sequences defined by Sloss (1963). Sequence 1 is approximately equivalent to Sloss's Sauk sequence, whereas most of the others are shorter than the subsequences which Sloss (1988) later defined. This suggests that the 10 sequences recognized in Ontario may actually be a combination of Mitchum et al.'s (1977) depositional sequences and Sloss's (1963) sequences. The 2 terms, sequence and depositional sequence, are used interchangeably in this chapter and Mitchum et al.'s (1977) definition is herein broadened to read: stratigraphic packages or intervals of genetically related sedimentary rocks which are commonly bounded by unconformities.

Chart D (Map 2582, map case) shows the location, age and duration of the stratigraphic sequences and provides the framework for this chapter. It does not show relative thickness of strata. The patterns on Chart D reflect the fact that subaerial exposure, and hence, erosional episodes punctuate the Phanerozoic record. Precisely how much material has been removed by erosion is somewhat conjectural; however, some evidence of the previous existence and original distribution of strata is available from the distribution and nature of Paleozoic and Mesozoic outliers (*see* Figures 20.1 and 20.2) on the Canadian Shield, from lithologic and biostratigraphic correlation, from indicators of depth of burial found in the preserved rock record, and from our understanding of the vertical movement of structural elements, such as arches and basins. Such evidence has enabled the construction of maps (*see* Figures 2 to 11, Map 2582, map case) showing the maximum (previous) and present (including the subsurface) distributions of the strata constituting the 10 sequences. The subcrop pattern (i.e., the areal distribution of strata occurring at surface and immediately below drift cover) of units comprising these 10 depositional sequences is shown on the regional bedrock geology maps (*see* Maps 2541 to 2544, map case) and the similar distribution of the sequences is shown on the tectonic maps (*see* Maps 2575 to 2578, map case).

## Regional Tectonic Setting

The Paleozoic and Mesozoic strata of Ontario occur in parts of 2 general areas (*see* Figure 20.1; Baer et al. 1977; Sanford and Baer 1981): the St. Lawrence Platform (which extends from the Great Lakes of Ontario and the United States to western Newfoundland) and the Hudson Platform. These platforms, in a regional sense, consist of relatively flat-lying and undeformed sedimentary rocks (Sanford et al. 1979), as compared to the more structurally complex terrain of the directly underlying Precambrian rocks of the Canadian Shield. These platforms have had a major influence on the physiographic development of the present day terrain, producing relatively flat, commonly swampy, lowlands.

The St. Lawrence Platform is subdivided, by basement uplifts, into 3 segments: the Eastern, Central and Western St. Lawrence platforms. Only parts of the central and western segments are present in Ontario (*see* Figure 20.1). The Central St. Lawrence Platform coincides roughly with the Ottawa–St. Lawrence Lowland physiographic region, whereas the Western St. Lawrence Platform includes the landmass assigned to the Great Lakes–St. Lawrence Lowlands physiographic region (*see* Figure 20.1; Chapman and Putnam 1984).

On the central Canadian Shield, Paleozoic and Mesozoic rocks of the Hudson Platform underlie Hudson Bay and James Bay, and extend into onshore areas of northern Ontario and adjacent areas of Quebec and Manitoba. The on land part of this platform is known as the Hudson Bay and James Bay lowlands physiographic regions (*see* Figure 20.1).

Typically, platformal areas are centred on or contain one or more depocentres, or basins, which are separated from adjacent basins by arches, that is, positive, roughly linear zones which have experienced periodic relative uplift. The basins and arches that controlled the distribution and

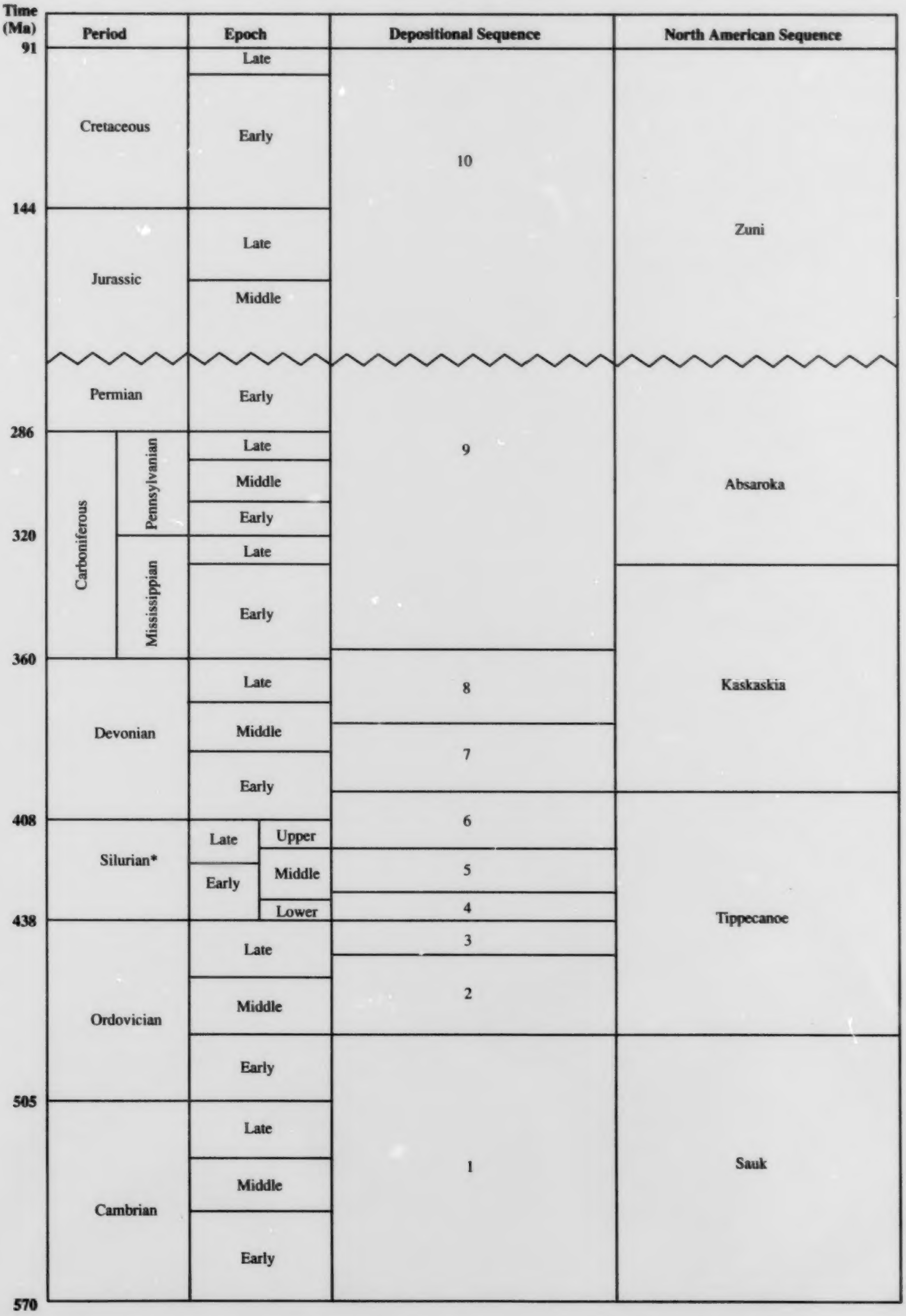
Time (Ma)
Period
Epoch
Depositional Sequence
North American Sequence
91
Late
Cretaceous
Early
10
144
Zuni
Late
Jurassic
Middle
Permian
Early
286
Late
9
Pennsylvanian
Middle
Absaroka
Early
Carboniferous
320
Late
Mississippian
Early
360
Late
8
Kaskaskia
Devonian
Middle
7
Early
408
6
Late
Upper
Silurian*
5
Early
Middle
4
Lower
438
Tippecanoe
3
Late
2
Ordovician
Middle
Early
505
Late
1
Sauk
Middle
Cambrian
Early
570

**Figure 20.3.** Age ranges of depositional sequences in Ontario, as compared with the corresponding North American sequences of Sloss (1963, 1988). *Both standard time-stratigraphic and North American rock-stratigraphic subdivisions shown for Silurian.

accumulation of sediments in Ontario during the Paleozoic and Mesozoic are shown in Figure 20.2.

The physiographic setting and major structural features controlling the rock distribution and character in each of the platform areas are discussed below.

## The Central St. Lawrence Platform

The Central St. Lawrence Platform (*see* Figure 20.1) is bounded by exposed Precambrian highlands (part of the Grenville Province of the Canadian Shield) to the north and west, the latter being the Frontenac Arch (or Axis) which trends southeast across eastern Ontario and into the Adirondack Mountains of New York State. The Ontario and adjacent New York State part of the Central St. Lawrence Platform is referred to as the Ottawa Embayment, the Ontario part of which covers approximately 10 000 km$^2$ and is underlain by sedimentary rocks ranging in age from Early(?) Cambrian to Late Ordovician (Wilson 1946; Baer et al. 1977; Wade et al. 1977; Williams 1991). These strata are laterally continuous with rocks of the Quebec Basin, which extends east and south of Montreal and is bounded on the south by the Appalachian structural front (*see* Figure 20.2; Clark 1972; Hofmann 1972; Riva 1972).

Unlike the more strongly folded, lithologically diverse strata of the Appalachian Basin lying to the southeast, which were deposited and subsequently deformed in a more tectonically active, continental margin setting, the strata in the Central St. Lawrence Platform were deposited on a relatively tectonically stable, shallow shelf, which was largely unaffected by Paleozoic orogenic events. These rocks are, as a result, generally flat lying and of comparatively uniform composition. During the Late Mesozoic, however, the opening of the Atlantic Ocean produced block faulting along the Ottawa–Bonnechere Graben (*see* Figure 20.2). This graben essentially parallels the original depositional embayment (i.e., the Ottawa Embayment), although block faulting of the Ordovician strata during the development of the graben was most severe in the northern part of the Central St. Lawrence Platform, along the Ottawa River.

Recent mapping by the Ontario Geological Survey (OGS) in eastern Ontario (Williams and Wolf 1982; Williams and Rae 1983; Williams 1991) has updated the stratigraphic knowledge of the up to 1130 m thick Ordovician succession and the relatively complex structural patterns in this area. Bedding is generally subhorizontal, except in close proximity to fault zones (Figure 20.4). Within these fault zones, steeply dipping normal faults with up to 1000 m of vertical displacement have been identified (Williams 1991).

## The Western St. Lawrence Platform

West of the Frontenac Arch and south of the Canadian Shield, Upper Cambrian to Upper Devonian (and possibly Lower Mississippian) strata underlie approximately 70 000 km$^2$ of southern Ontario, forming part of the Western St. Lawrence Platform (*see* Figure 20.1). Ordovician and Silurian strata exposed on Manitoulin Island, Cockburn Island and St. Joseph Island (the North Channel Islands) in northern Lake Huron are also included within this platform.

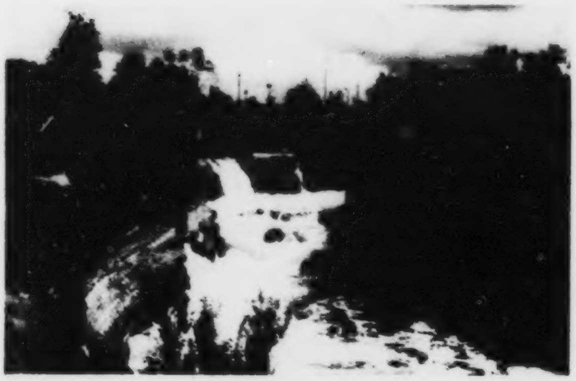

**Figure 20.4.** Faulted anticline in the Middle Ordovician (Sequence 2) Rockcliffe and Shadow Lake formations at Prince of Wales Falls, Hog's Back Park, Ottawa. (Photo: DA89–12–20)

The stratigraphic succession of the Western St. Lawrence Platform consists of a variety of sediment types which were deposited mainly in shallow epicontinental seas. Superficially, these rocks are gently dipping (toward basin centres) and display little evidence of structural disturbance, although, deposition of these units actually took place in and around 3 major paleogeographic structural elements (i.e., basins and arches; *see* Figure 20.2):

a) Appalachian Basin: a northeast-trending, elongate, foreland basin that extends some 1400 km from the Great Lakes area to Alabama. This structure has a maximum width of about 800 km; the strata filling the basin attain a maximum thickness of about 13 000 m in the northern and eastern parts of it (Milici and de Witt 1988). As a result of intermittent orogenic activity along the eastern margin of this basin throughout the Paleozoic, the deposits within this basin are mostly clastic dominated (i.e., siliciclastic or terrigenous).

b) Michigan Basin: a roughly circular intracratonic basin, roughly 650 km in diameter, centred on the southern peninsula of Michigan. The Paleozoic and Mesozoic succession in the basin reaches a maximum thickness of about 4800 m (Fisher et al. 1988) and consists predominantly of carbonate rocks due to its relatively distant position from the Appalachian orogenic belt.

c) Algonquin Arch: a broad basement ridge forming the spine of the southwestern Ontario peninsula. The arch is a northeastern extension of the Findlay Arch, with the Chatham Sag (located between Lake Huron and Lake Erie) separating the 2 arches. These 2 arches acted as a hinge line between the Appalachian Basin and Michigan Basin during the Paleozoic and they periodically moved vertically, and probably shifted laterally, in response to tectonism at the craton margin to the southeast (Quinlan and Beaumont 1984).

North of the Algonquin Arch, strata generally exhibit consistent regional dips of 6 to 9 m/km westward into the

Michigan Basin, whereas south of the arch the strata generally have dips of 6 m/km southward into the Appalachian Basin (Winder and Sanford 1972). Variations exist around the noses of the arches and in the Chatham Sag, where strata are generally flat lying. The interbasinal position of southwestern Ontario results in the maximum preserved thickness of only 1525 m of Paleozoic sedimentary rocks (Winder and Sanford 1972), as compared to the maximum preserved thicknesses of sedimentary rocks in the Michigan and Appalachian basins of 4800 m and 13 000 m, respectively.

Recent regional-scale, deep drilling investigations by the Ontario Geological Survey (Johnson 1983; Johnson et al. 1983b, 1985; Wolf 1986) have added much to our understanding of the Paleozoic stratigraphy of the Western St. Lawrence Platform. The locations of the 11 drill holes, which were cored from Paleozoic bedrock surface to Precambrian basement, are shown in Figure 20.5.

## The Hudson Platform

The Hudson Platform (*see* Figure 20.1) covers 1.2 million km², with approximately 240 000 km² underlying the Hudson Bay and James Bay lowlands in northern Ontario (Sanford 1987). The platform, bounded by Precambrian uplands, includes the Moose River and Hudson Bay basins (*see* Figure 20.2). The uplands include the northeast-trending Fraserdale Arch and Cape Henrietta Maria Arch. The latter, probably part of the Transcontinental Arch, separates the Hudson Bay Basin and the Moose River Basin (*see* Figure 20.2). Bounding the platform on the southwest is the Severn Arch, a positive northwest-trending feature that was probably continuous with the Frontenac Arch to the south.

The Hudson Bay Basin, centred in Hudson Bay, is a roughly circular intracratonic basin, approximately 1100 km in length (north trending) and 850 km in width, with only a small part on shore. Ordovician to Devonian and Cretaceous strata dip gently towards the basin centre, where the sedimentary succession attains a maximum thickness of about 2000 m (Sanford 1987; Sanford and Grant 1990). At the western shoreline of Hudson Bay, thicknesses of approximately 1000 m of sediment are known (Thorpe 1988).

In the extreme southern part of the Hudson Platform, Ordovician to Devonian, Middle Jurassic and Lower Cretaceous strata dip gently into the small intracratonic Moose River Basin (roughly 400 by 250 km in size), attaining a composite thickness of about 800 m. The southern margin of the Moose River Basin is, in part, fault bounded and displays a number of smaller-scale structural features, the most significant being the Pivabiska Ridge, Grand Rapids High and Moose River High (Figure 20.6; Sanford and Norris 1975; Bezys 1990). Although relatively small, these structural highs had, at certain times, affected the local sedimentation pattern (i.e., thickness and facies) of the various units in the area. The strata display gentle dips on the northern flank of the basin, and somewhat steeper dips on the partly faulted southern flank (Sanford and Norris 1973). Faulting of Cretaceous sediments within this basin indicates that tectonic movements continued in this platform after this time.

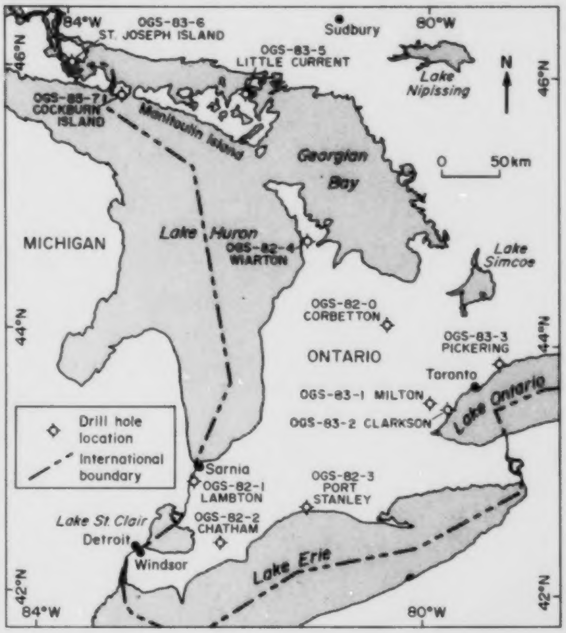

**Figure 20.5.** Ontario Geological Survey regional deep drill hole locations in southern Ontario.

## DEPOSITIONAL SEQUENCE 1: EARLY CAMBRIAN (WAUCOBAN) TO EARLY ORDOVICIAN (CANADIAN)

### Overview

Sequence 1 is dominated by nonmarine, craton-derived clastic rocks which are succeeded by carbonate rocks. The sequence is bounded by a basal erosional contact with the underlying Precambrian Shield, and an upper erosional surface brought about by Middle Ordovician uplift related to the onset of the Taconic Orogeny.

The initial deposits of Sequence 1 are nonmarine sandstones, shales and conglomerates derived from uplifted regional-scale arches in the Precambrian Shield (Poole et al. 1970; *see* Figure 20.2). These sediments were deposited on a faulted basement surface formed by rifting during initial continental breakup and separation (James and Stevens 1982; Cumming 1983; Williams and Smyth 1983). Following deposition, these beds were tilted locally by block faulting, and were subjected to widespread erosion during the late Early and Middle Cambrian.

In the Central and Eastern St. Lawrence platforms (*see* Figure 20.1), the basal units of Sequence 1 are tentatively assigned a Neoproterozoic to Early Cambrian age, based on the age determination of correlative rocks in western

Newfoundland and Labrador (Cumming 1983). Possible remnants of this deposition in Ontario (Covey Hill Formation, Potsdam Group) occur in the Ottawa Embayment and on the flanks of the Frontenac Arch (*see* Figure 20.2; Clark 1972; Williams 1991).

In the Sault Ste. Marie area, sandstones of the Jacobsville Formation (not shown on Chart D, Map 2582, map case) may also represent initial Sequence 1 sedimentation (Hamblin 1958; Russell 1987; Sanford, in press). However, paleomagnetic evidence indicates a Keweenawan age for the Jacobsville Formation (Roy and Robertson 1978), and thus, this unit is discussed in the Proterozoic chapter of this volume (*see* Sutcliffe, this volume). There is no evidence for similar deposits in other parts of Ontario.

With the opening of Iapetus Ocean (proto–Atlantic) in the mid–Early(?) Cambrian, craton-derived clastic sediments (mainly orthoquartzites) and carbonate sediments were deposited on a broad continental shelf (prior to Appalachian Basin development) which was open to the ocean to the southeast (Figure 2, Map 2582, map case; James and Stevens 1982; Cumming 1983; Sanford, in press).

Although Lower and Middle Cambrian strata were deposited over the Eastern St. Lawrence Platform (i.e., in the Quebec Basin; Cumming 1983), it was not until the Late Cambrian that a seaway transgressed into western Quebec (Clark 1972), eastern Ontario (Wilson 1946; Williams 1991) and southwestern Ontario and the adjacent United States, thereby initiating deposition within the Michigan Basin (*see* Figure 20.2; Cohee 1945; Hamblin 1958; Flagler 1966). Initial deposits of this transgression in eastern Ontario consist of craton-derived orthoquartzites and conglomerates (Nepean Formation, Potsdam Group; Wilson 1946; Williams 1991). These deposits grade upward into Lower Ordovician dolostones (Beekmantown Group).

In southwestern Ontario, initial deposits from the transgression were orthoquartzites, succeeded gradationally by dolostones that onlapped and ultimately overlapped the Algonquin Arch. These deposits are subdivided into several formations known only in the subsurface (Sanford and Quillian 1959; Rickard 1973; Bailey and Cochrane 1984a). It is speculated that deposition continued in southwestern Ontario through to the Early Ordovician, with deposition of strata equivalent to the Lower Ordovician units of the

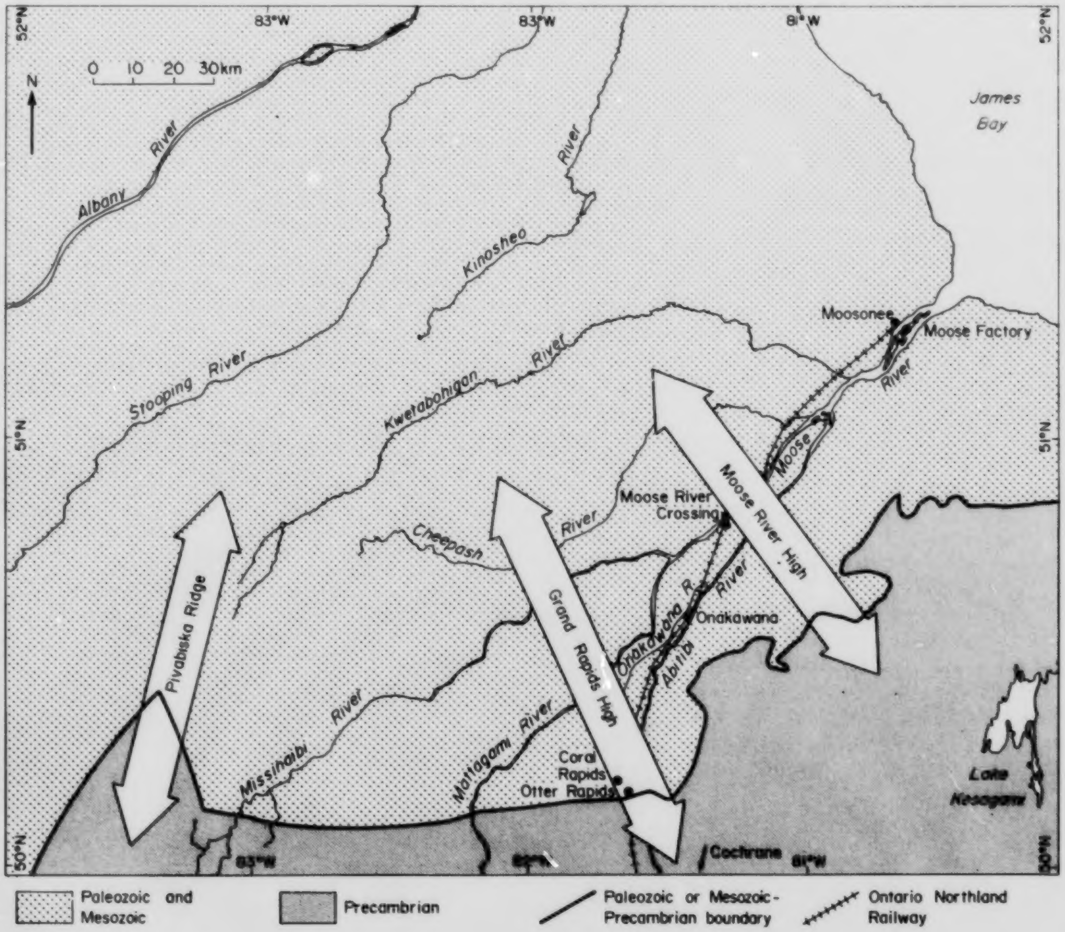
52°N
83°W
83°W
81°W
52°N
0 10 20 30km
N
James Bay
Albany River
River
Kinosheo
Stooping River
Kwetabohigan River
Moosonee
Moose Factory
Moose River
51°N
51°N
Moose River High
Moose River Crossing
River
Cheepash
Pivabiska Ridge
Grand Rapids High
Onakawana R.
Abitibi River
Onakawana
Missinaibi River
Mattagami River
Coral Rapids
Otter Rapids
Lake Kesagami
50°N
83°W
82°W
Cochrane
81°W
50°N
Paleozoic and Mesozoic
Precambrian
Paleozoic or Mesozoic-Precambrian boundary
Ontario Northland Railway

**Figure 20.6.** Structural elements in the southern part of the Moose River Basin (*modified from* Bezys 1990).

Michigan and Appalachian basins; thus, such deposits may have once blanketed much of southern Ontario.

Widespread erosion during the Middle Ordovician stripped off most of this sequence in Ontario. This erosion resulted from uplift triggered by the initiation of convergence along the eastern margin of the craton during the early Middle Ordovician (Whiterockian) (James and Stevens 1982; Cumming 1983).

The amount of relief on the Precambrian terrain over which the earliest Paleozoic transgressions occurred is not well known. Based on the present relief of the shield and the thickness of Paleozoic strata required to blanket the Precambrian, in eastern Ontario, the total relief may have been in excess of 100 m (Williams 1991). This relief is not apparent on the regional structure contour map of the Precambrian due to the scarcity of data (Bailey and Cochrane 1984a; *see also* Figure 20.2). As land plants had not yet evolved, the land surface would have been a landscape of rocky promontories and locally derived regolith.

For lithological details of the units comprising Sequence 1, the reader is referred to Table 20.1. The present subcrop and outcrop distribution of these units is shown on Maps 2541 to 2544 (map case). The subcrop and outcrop distribution of the sequence as a whole is shown on Maps 2575 to 2578 (map case). The pre-erosion distribution of Sequence 1 is shown on Figure 2, Map 2582 (map case). Figure 20.7 is a stratigraphic column showing the units of Sequence 1. In the following section, details of the stratigraphy of Sequence 1 are presented, subdivided on a geographic basis.

## Geology

### EASTERN ONTARIO

The oldest Paleozoic sedimentary rocks in Ontario are considered to be the nonmarine clastic rocks of the Covey Hill Formation (Potsdam Group) of the Central St. Lawrence Platform (Clark 1966, 1972; Williams and Wolf 1982; Williams 1991). The Covey Hill Formation, estimated to be Early to Late Cambrian, consists of a relatively thin succession of feldspathic conglomerates and sandstones. This poorly exposed unit only outcrops in the southwestern part of eastern Ontario (Figure 20.8; Map 2544; Figure 2, Map 2582, map case) and its distribution is limited to depressions on the Precambrian surface (Williams 1991). Rare exposures of the basal contact of this unit with the Precambrian commonly display rounded cobbles of the underlying Precambrian rock (Figure 20.9; Williams 1991). An alluvial fan and braided fluvial depositional environment has been suggested for this unit (Wolf and Dalrymple 1984, 1985).

Strata of the overlying Nepean Formation (Potsdam Group) contain the earliest marine Paleozoic sedimentary rocks in eastern Ontario and consist of a thick succession of quartz-rich sandstones and minor conglomerates (Wilson 1946; Williams 1991). Its basal, unconformable contact with the Covey Hill Formation is marked by the first appearance of feldspar-free sandstone; its upper conformable contact with the Beekmantown Group is marked by the lowermost dolomitic bed. Within this unit are unique vertical, cylindrical sandstone "pillars" (Figure 20.10) with horizontal and vertical dimensions of up to 5 m (Williams 1991). Of the many theories proposed to explain the origin of these enigmatic structures (Hawley and Hart 1934; Dietrich 1953), theories involving dewatering seem most likely. Biota in the upper part of the Nepean Formation and overlying March Formation indicate a Late Cambrian to Early Ordovician age for the Nepean Formation (Greggs and Bond 1971; Bond and Greggs 1973; Brand and Rust 1977). Most of the Nepean Formation is interpreted as subtidal to intertidal in origin, although terrestrial facies (e.g., coastal dunes and fluvial channels) are also present (Wolf and Dalrymple 1984, 1985).

Based on rock type and stratigraphic position, the Nepean Formation of eastern Ontario is approximately stratigraphically equivalent to at least the basal part of the Potsdam and Mount Simon formations (described below) or possibly the entire Cambrian interval in the subsurface of southwestern Ontario (*see* Figure 20.7; Chart D, Map 2582, map case). There is little doubt that these Cambrian units are diachronous, but a lack of detailed paleontology precludes detailed correlation in southern Ontario.

The only Lower Ordovician deposits in Ontario are the shallow marine carbonate and clastic rocks of the Beekmantown Group, which is subdivided into the March and Oxford formations (Williams 1991; Chart D, Map 2582, map case; *see* Figure 20.7). This group succeeds the Potsdam Group and is restricted to eastern Ontario (with minor remnants on the west side of the Frontenac Arch; *see* Figure 20.8; Williams 1991). The Beekmantown Group has been assigned to the Lower Ordovician based on conodonts of early Canadian (March) to late Canadian (Oxford) age and on upper Early Ordovician trilobites found in the Oxford Formation (Greggs and Bond 1971; Bond and Greggs 1973; Ludvigsen 1979a).

The March Formation consists of interbedded sandstones and dolostones (Wilson 1946; Bond and Greggs 1973; Williams 1991). The lower contact of the March Formation with the Nepean Formation is conformable. Locally, the March Formation rests unconformably on Precambrian knobs which protrude through the Nepean Formation; in these areas, the unit is characterized by boulder- and cobble-sized intraclasts of quartzite. The upper contact of the March Formation with the Oxford Formation is conformable and arbitrarily placed at the last occurrence of sand. The presence of calcite-filled vugs and algal lamination in the dolostones, and bipolar cross-bedding in the sandstones, suggests depositional environments ranging from supratidal to subtidal (Williams 1991).

Dolostones of the overlying Oxford Formation contain algal laminae, quartz- and calcite-filled vugs, desiccation cracks, stylolites, burrows, soft-sediment deformation structures and rare interbeds of glauconitic shale (Wilson 1946). Based on these features, Williams (1991) has suggested a supratidal to intertidal, hypersaline depositional environment for this unit.

**Table 20.1.** Early Cambrian (Waucoban) to Early Ordovician (Canadian) units in Ontario—Depositional Sequence 1 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Eastern Ontario** | | | | | | |
| Oxford Formation (Beekmantown Group) | Eastern Ontario and minor remnants on west side of Frontenac Arch | Canadian | Supratidal to intertidal, hypersaline | up to 200 m | Brown- to green-grey, sublithographic to medium-crystalline dolostone; thin to thick bedded; occasional interbeds of glauconitic shale (up to 30 cm thick); fine- to medium-grained quartz sandstone and shaly dolostone interbeds common in basal part of unit.<br>Algal laminae; quartz- and calcite-filled vugs; desiccation cracks; stylolites; burrows; soft-sediment deformation structures.<br>Upper contact unconformable; conformable lower contact with the March Fm. is arbitrarily drawn at the upper limit of common occurrence of quartz sand. | Williams 1991;<br>Wilson 1946 |
| March Formation (Beekmantown Group) | Eastern Ontario and minor remnants on west side of Frontenac Arch | Canadian | Supratidal to subtidal | 6 to 64 m | Interbedded quartz sandstone, dolomitic sandstone, sandy dolostone and dolostone.<br>Sandstones are grey, brown and green, fine to coarse grained and well sorted.<br>Sandy dolostones are brown- to blue-grey and thin to thick bedded.<br>Dolostones are brown- to green-grey, thin to thick bedded and micro- to medium-crystalline.<br>Calcite-filled vugs and algal laminae in dolostones; bipolar cross-bedding in sandstones.<br>Lower contact is conformable with Nepean Fm. and is defined at the base of the lowest dolomitic bed; locally unconformable on Precambrian knobs, where boulder- to cobble-sized quartzite clasts are present in the March Fm. | Wilson 1946;<br>Bond and Greggs 1973;<br>Williams 1991 |
| Nepean Formation (Potsdam Group) | Eastern Ontario | Croixian | Mostly subtidal to intertidal; however, terrestrial deposition (coastal dunes and fluvial channels) is also indicated | up to 300 m | White, light grey, brown or green quartz-rich sandstone with conglomerate interbeds; thin bedded to massive.<br>Cross-bedding in sandstone; vertical, cylindrical sandstone pillars, up to ~5 m in diameter by 5 m high.<br>Lower contact with the Covey Hill Fm. is unconformable and is defined by the lowest occurrence of feldspar-free sandstone. | Wilson 1946;<br>Williams 1991;<br>Greggs and Bond 1971;<br>Bond and Greggs 1973;<br>Brand and Rust 1977 |

Table 20.1. Early Cambrian (Waucoban) to Early Ordovician (Canadian) units in Ontario—Depositional Sequence 1 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Covey Hill Formation (Potsdam Group) | Eastern Ontario | Waucoban to Albertan(?) | Alluvial fan and braided fluvial environment | up to 13 m | Very poorly sorted, feldspathic conglomerate and sandstone.<br>Conglomerates are light grey, dark grey to red-brown and poorly stratified to massive; contain well-rounded, pebble- to cobble-sized Precambrian clasts where in contact with Precambrian basement.<br>Sandstones are thin to thick bedded and poor to well sorted.<br>Large-scale (up to 1 m) cross-bedding in sandstone.<br>Lower contact rests unconformably on Precambrian basement. | Williams and Wolf 1982; Williams 1991; Clark 1966, 1972; Wolf and Dalrymple 1984, 1985 |
| **Southwestern and South-Central Ontario East of Longitude 81° W** | | | | | | |
| Little Falls Formation | Southwestern Ontario (subsurface) | Croixian | Undetermined | up to 31 m | Light brown, fine- to medium-crystalline dolostone with oolitic stringers.<br>Lower contact with Theresa Fm. is gradational. | Sanford and Quillian 1959; Winder and Sanford 1972 |
| Theresa Formation | Southwestern Ontario (subsurface) | Croixian | Undetermined | up to 107 m | Orthoquartzitic sandstone interbedded with buff and grey-buff, fine-crystalline, oolitic dolostone and sandy dolostone. In Niagara Peninsula, upper part of unit contains small concentrations of glauconite, and red shaly dolostone and shale are locally present at base of unit.<br>Lower contact is conformable with the Potsdam Fm. | Sanford and Quillian 1959; Winder and Sanford 1972 |
| Potsdam Formation | Southwestern Ontario (subsurface) | Croixian | Undetermined | up to 46 m | Arkosic conglomerate and red, white and grey orthoquartzitic sandstone.<br>Basal, medium-grained quartzose sandstone occurs at the contact with the underlying Precambrian rocks.<br>Lower contact is unconformable with Precambrian. | Winder and Sanford 1972; Sanford and Quillian 1959 |
| **West of Longitude 81° W** | | | | | | |
| Prairie du Chien Group | Northern Lake Huron (subsurface) | Canadian | Undetermined | up to 31 m | Sandy dolostone.<br>Unconformable upper contact. Lower contact unconformable with Precambrian, or conformable(?) with Trempealeau Fm. | Wolf 1986; Dorr and Eschman 1970 |

**Table 20.1.** Early Cambrian (Waucoban) to Early Ordovician (Canadian) units in Ontario—Depositional Sequence 1 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Trempealeau Formation | Southwestern Ontario (subsurface) | Croixian | Undetermined | up to 75 m | Buff to grey-buff, fine- to medium-crystalline dolostone containing oolitic stringers. Locally up to 6 m of orthoquartzitic sandstone, with well-rounded and frosted grains, at top of unit.<br>Locally pink dolostone near base. Base consists of 1 to 2 m thick arkosic sandstone where in contact with Precambrian. Lower contact conformable(?) with Eau Claire Fm. and unconformable with Mount Simon Fm. or Precambrian. | Sanford and Quillian 1959 |
| Eau Claire Formation | Southwestern Ontario (subsurface) | Croixian | Upper unit: low energy tidal flat<br>Lower unit: oolitic shoal formed in shallow, agitated environment | up to 80 m | Upper unit is light to medium brown, fine- to coarse-grained, dolomitic sandstone with rare to abundant, grey to green shale partings; glauconite grains scattered throughout; cross-bedding; planar laminae.<br>Lower unit is light to medium grey-brown, sandy, oolitic dolostone, commonly bioturbated and stylolitic.<br>Lower contact conformable and gradational with Mount Simon Fm.. Basal arkose (1 to 2 m thick) is present where in contact (unconformable) with Precambrian. | Sanford and Quillian 1959; Trevail 1990 |
| Mount Simon Formation | Southwestern Ontario (subsurface) | Croixian | Braided fluvial deposits reworked in tidal flat-tidal channel environment | up to 50 m | Grey-brown, feldspathic sandstone; feldspar content (30% max.) gives rock a pink cast.<br>White to light grey quartzose sandstone having rounded grains at top of unit and angular grains at base. Cross-laminae; ripple cross-laminae; parallel laminae.<br>Basal conglomerate and arkose may be present locally.<br>Lower contact is unconformable with Precambrian. | Trevail 1990; Sanford and Quillian 1959 |
| Munising Formation | St. Joseph and Campement d'Ours Islands,<br>Sault Ste. Marie area | Croixian(?) | Shoreline (eolian) to shallow marine shelf environment (in Michigan) | up to 70 m | Bioturbated, sandy dolostone and siltstone, laminated siltstone, and poorly sorted quartz conglomerate; minor interbedded, 10 to 20 cm thick, well-sorted sandstone and dark grey and green mudstone.<br>Lower contact is unconformable on the Precambrian (Jacobsville sandstone or the Canadian Shield). | Russell 1985; Hamblin 1958; Haddox and Dott 1990 |

Early Ordovician deposits equivalent to the Beekmantown Group may have extended across southern Ontario and been continuous with the Prairie du Chien Group of the Michigan Basin (described below; Figure 2 and Chart D, Map 2582, map case). These deposits were removed by erosion at the close of Sequence 1, resulting in their much reduced present distribution.

## SOUTHWESTERN AND SOUTH-CENTRAL ONTARIO

In southwestern and south-central Ontario, the initial Cambrian marine transgression is represented by sandstones succeeded gradationally by dolostones that progressively onlapped and ultimately overlapped the Algonquin and Findlay arches (Figure 2 and Chart D, Map 2582, map case). These "basal beds" (Caley 1943; Sanford and Quillian 1959) were subsequently eroded from the arches and overlain unconformably by Middle Ordovician clastic and carbonate units of Sequence 2 (Beards 1967; Liberty 1969; Bailey and Cochrane 1984a). The present distribution of these units around the Algonquin Arch approximately follows the outline of the southwestern Ontario peninsula, suggesting that the arch configuration during the Cambrian to Early Ordovician was much the same as at present (Sanford and Quillian 1959).

From their pinch-out along the southeastern flank of the Algonquin Arch (*see* Figure 20.2), the Cambrian units increase in thickness beneath Lake Erie, reaching a

| **Series** | **Michigan Basin** | | **Appalachian Basin** | | |
|---|---|---|---|---|---|
| | **Southwestern and South-Central Ontario** | | | **Eastern Ontario** | |
| Lower Ordovician | Prairie du Chien Gp. | | | Beekmantown Gp. | Oxford Fm. |
| | | | | | March Fm. |
| | ? | ? | ? | ? | |
| Upper Cambrian | Munising Fm. | Trempealeau Fm. | Little Falls Fm. | Nepean Fm.<br>(Potsdam Gp.) | |
| | | Eau Claire Fm. | Theresa Fm. | | |
| | | Mount Simon Fm. | Potsdam Fm. | | |
| | ? | ? | ? | ? | |
| Middle Cambrian | | | | | |
| | | | | ? | |
| | | | | Covey Hill Fm.<br>(Potsdam Gp.) | |
| Lower Cambrian | | | | | |
| | | | | ? | |

**Figure 20.7.** Stratigraphic units of Sequence 1 (? indicates uncertain age).

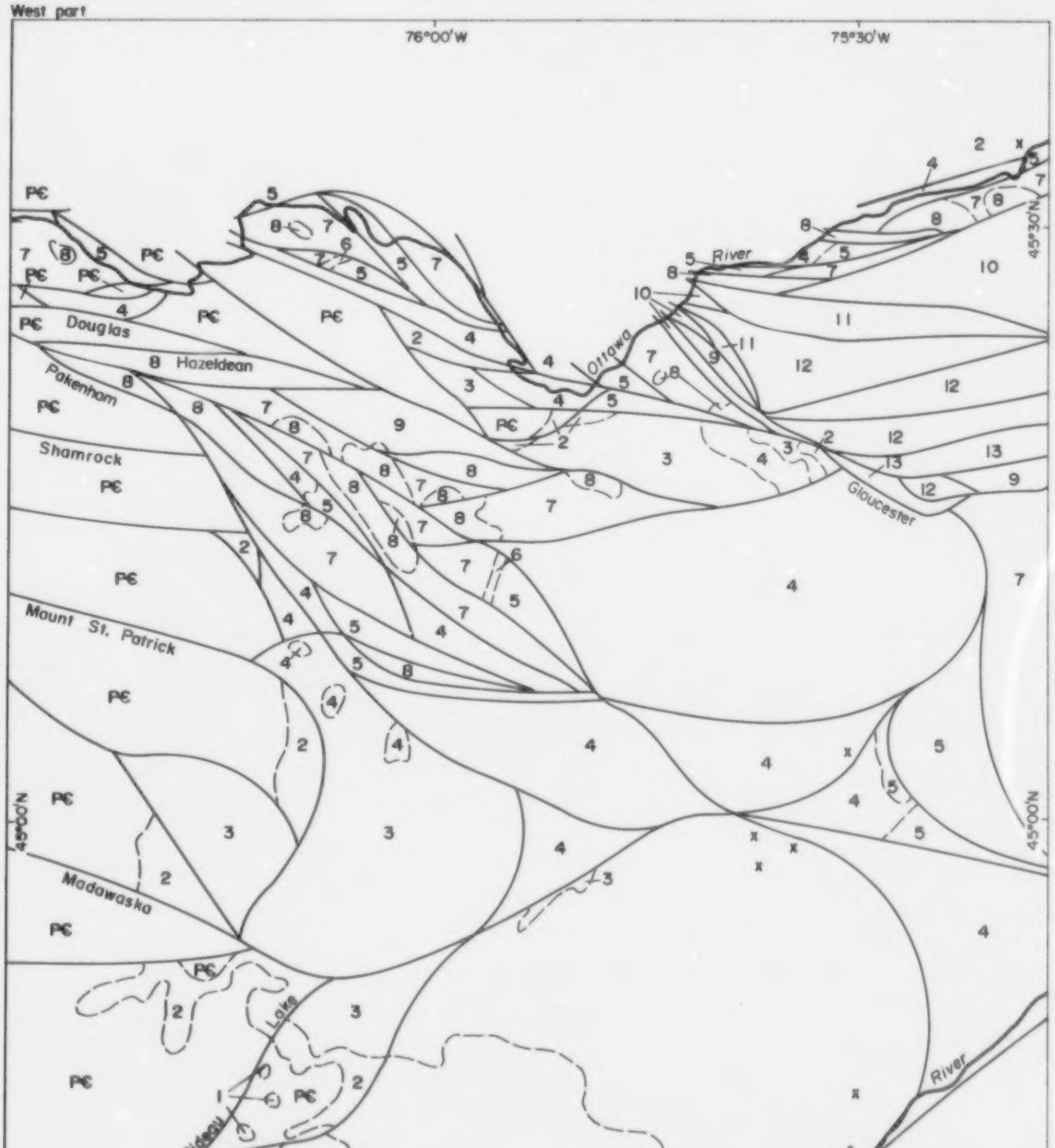
West part
76°00'W
75°30'W
45°30'N
45°00'N
44°30'N
Douglas
Hazeldean
Pakenham
Shamrock
Mount St. Patrick
Madawaska
Ottawa
River
Gloucester
Rideau
Lake
St. Lawrence
River
PЄ
1
2
3
4
5
6
7
8
9
10
11
12
13
4-7
x

**Figure 20.8.** Generalized Paleozoic geology of eastern Ontario (*modified from* Williams 1991).

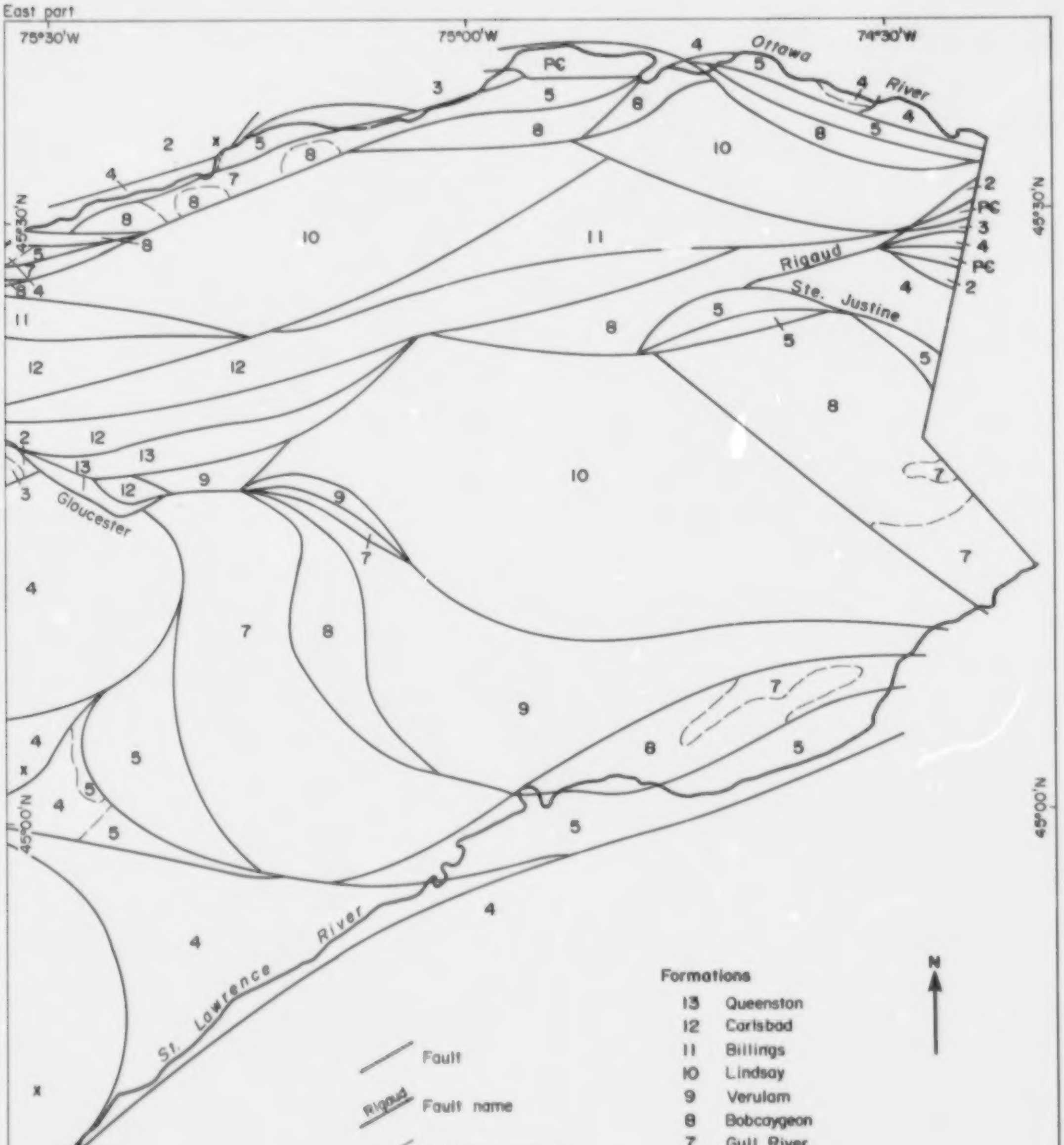
East part
75°30'W
75°00'W
74°30'W
45°30'N
45°00'N
44°30'N
Ottawa River
Rigaud
Ste. Justine
Gloucester
St. Lawrence River
PC
Fault
Fault name
Geological contact
x Precambrian Inlier
0
20 km
N
Formations
13 Queenston
12 Carlsbad
11 Billings
10 Lindsay
9 Verulam
8 Bobcaygeon
7 Gull River
6 Shadow Lake
5 Rockcliffe
4 Oxford
3 March
2 Nepean
1 Covey Hill
PC Precambrian

**Figure 20.8.** Generalized Paleozoic geology of eastern Ontario (*modified from* Williams 1991).

**Figure 20.9.** Conglomerate and sandstone of the Covey Hill Formation, Mill Pond Quarry, south of Perth (*courtesy of* Williams 1991). (Photo: RW82–3–34)

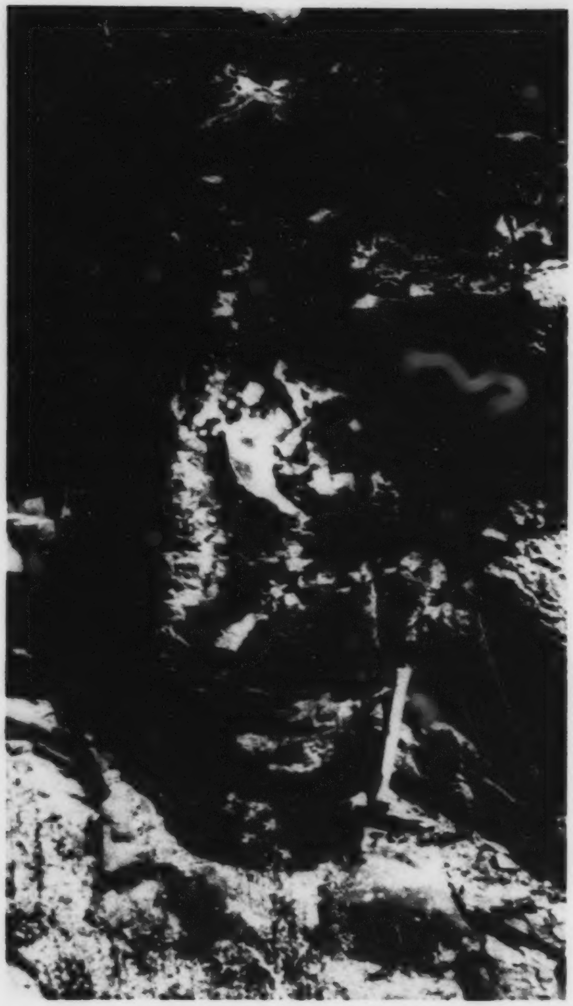

**Figure 20.10.** Cylindrical sandstone structure, or "pillar", in quartz sandstone of the Nepean Formation, Highway 15 roadcut, north of Morton (*courtesy of* Williams 1991). (Photo: RW82–4–14)

maximum of 2170 m in the central Appalachian Basin (Winder and Sanford 1972). On the northwestern side of the arch, the strata are at most 77 m thick in Ontario, thickening westward to 620 m in the south-central part of the Michigan Basin (Winder and Sanford 1972).

The combination of Appalachian and Michigan Basin stratigraphic terminology used for these Cambrian "basal beds" is listed in Figure 20.7. The dividing line between the 2 schemes is arbitrarily drawn at long. 81°W (Sanford and Quillian 1959).

## EAST OF LONGITUDE 81°W

The Potsdam Formation consists of arkosic conglomerate and orthoquartzitic sandstone (Winder and Sanford 1972). The conformably overlying Theresa Formation consists of orthoquartzitic sandstone with interbeds of dolostone (Sanford and Quillian 1959; Winder and Sanford 1972). Minor glauconite occurs throughout the uppermost part of the Theresa Formation and, in the Niagara Peninsula, shaly dolostone and shale are present locally at the base of this unit (Sanford and Quillian 1959). The overlying Little Falls Formation consists of dolostones (Winder and Sanford 1972).

Rickard (1973) noted that the Theresa Formation of western New York (and southwestern Ontario) is not continuous with the Theresa Formation in north-central New York. The latter unit is Ordovician and correlative with the March Formation (Beekmantown Group) of eastern Ontario (Williams 1991).

## WEST OF LONGITUDE 81°W

The Mount Simon Formation consists predominantly of feldspathic sandstone. Its relatively high feldspar content (up to 30%) may be attributed to weathering from the underlying granitic Precambrian rocks (Trevail 1990) or to diagenetic alteration (Longstaffe et al. 1990), and gives the rock a pinkish cast. The sedimentary structures suggest the Mount Simon Formation is a tidal flat–tidal channel deposit, with the detritus having been supplied by the erosion of Precambrian terrain by a braided river system (Trevail 1990).

The overlying Eau Claire Formation (Sanford and Quillian 1959) has been subdivided by Trevail (1990) into a lower oolitic dolostone unit and an upper dolomitic sandstone unit. The lower unit is commonly bioturbated and stylolitic and interpreted as an oolitic shoal deposit, formed in shallow, agitated waters seaward from the tidal flat deposits of the Mount Simon Formation (Trevail 1990). Cross-bedding and planar laminations are recognized in the upper dolomitic sandstones (Trevail 1990) and a low-energy tidal flat depositional environment has been suggested for this unit (Trevail 1990). The Eau Claire Formation reaches a maximum thickness in Ontario of about 75 m under western Lake Erie. In some areas, the Eau Claire Formation overlaps the Mount Simon Formation and rests directly on Precambrian basement; in these areas, a basal arkose (1 to 2 m thick) is present (Sanford and Quillian 1959).

Overlying the Eau Claire Formation are the dolostones of the Trempealeau Formation. In Essex and Kent counties, as much as 6 m of orthoquartzitic sandstone may occur at the top of the formation. Locally, where the Trempealeau Formation rests directly on the Mount Simon Formation or the Precambrian surface, the base of the unit consists of arkosic sandstone 1 to 2 m thick (Sanford and Quillian 1959).

In Michigan, sandy dolostones of the Lower Ordovician Prairie du Chien Group overlie the Trempealeau Formation (Dorr and Eschman 1970) and probably correlate with the Beekmantown Group (Chart D, Map 2582, map case; *see* Figure 20.7). The Prairie du Chien Group may also be present in Ontario, in the subsurface beneath Cockburn Island along the north shore of Lake Huron (Wolf 1986, in prep.).

At the far west end of the Western St. Lawrence Platform, on St. Joseph Island and Campement d'Ours Island near Sault Ste. Marie, the Upper Cambrian Munising Formation, possibly stratigraphically equivalent to (or perhaps younger than) the Eau Claire and Mount Simon formations (*see* Figure 20.7; Chart D, Map 2582, map case), forms the base of the Paleozoic succession (Hamblin 1958; Poole et al. 1970; Russell 1985; Haddox and Dott 1990). The Munising Formation is complex in Ontario, such that the 3 well-defined members (basal conglomerate, Chapel Rock and Miners Castle members) identified by Hamblin (1958) in northern Michigan cannot be distinguished (Russell 1985). The Munising Formation in Ontario consists of dolostones, siltstones and poorly sorted conglomerates (Russell 1985).

Hamblin (1958) interpreted the Munising Formation in Michigan to represent a shallow marine setting, in a small subbasin, north of and separated from the Michigan Basin by the Northern Michigan Highland. Haddox and Dott (1990) identified shoreline depositional environments in the Munising Formation (also in Michigan) ranging from eolian to marine shelf in a generally deepening upward succession.

No Cambrian or Lower Ordovician sedimentary rocks have been identified in the Hudson Platform, the oldest strata being the upper Middle Ordovician Bad Cache Rapids Group (*see* Figure 20.11, Sequence 2; Sanford 1987). The basal beds of the Bad Cache Rapids Group, however, are locally derived clastic rocks, and a part of these may have been deposited during Sequence 1 time.

## Economic Geology

Interest in the oil and gas potential of Cambro–Ordovician units of southwestern Ontario is increasing as shallower resources are being depleted. Most of the hydrocarbon traps in these rocks occur where sandstone strata pinch out against the southeastern flank of the Algonquin Arch (Bailey and Cochrane 1984a). The lateral boundaries of the traps are faults, with the cap rock being the less permeable Ordovician dolostone or the overlying sandstones of the Middle Ordovician Shadow Lake Formation (*see* Sequence 2). Pools of this type include Clearville, Gobles, Innerkip and Willey (Bailey and Cochrane 1984a; Trevail 1990).

Although depth of drilling makes exploration and development costs high, these traps produce high quality oil and will usually run for several years without the need for enhanced recovery methods. Low wax content of the oil and the proximity to market continue to make these strata attractive for exploration. In 1989, total oil production from Cambrian strata amounted to 6520 $m^3$, or 2.6% of the Ontario oil production (Carter 1990). Total gas production for this period from these same units was 19 043 thousand $m^3$, or 3.9% of Ontario gas production (Carter 1990). The Innerkip pool is the only major producer of natural gas, with a cumulative production to the end of 1989 of 112.2 million $m^3$ (Trevail 1990).

The Beekmantown Group is a significant source of skid-resistant aggregate and crushed stone for both road and specialized construction use in eastern and south-central Ontario (Bezys and Johnson 1988; Derry, Michener, Booth and Wahl and OGS 1989b). The Nepean Formation and March Formation sandstones have been significant sources of building and dimension stone (Verschuren et al. 1985; Williams 1991). The Nepean Formation is currently exploited as a source of silica sand (Powell and Klugman 1979; Williams 1991).

# DEPOSITIONAL SEQUENCE 2: MIDDLE ORDOVICIAN (WHITEROCKIAN) TO EARLY LATE ORDOVICIAN (MAYSVILLIAN)

## Overview

Sequence 2 is bounded by a basal erosional contact resulting from an early Middle Ordovician regression. This boundary is roughly equivalent to the Sauk–Tippecanoe sequence boundary of Sloss (1963). The sequence consists of initial craton-derived clastic rocks succeeded by shallow water carbonates with minor craton-derived and orogen-derived clastic rocks. The upper boundary of this sequence is locally erosional and marked by a significant influx of orogen-derived clastic sediments.

The early Middle Ordovician (Whiterockian) regression of inland seas from the North American continent resulted in subaerial exposure and erosion of much of Sequence 1. Erosion appears to have continued unabated until Blackriverian time in most of Ontario and the surrounding area. However, in eastern Ontario, a small pulse of craton-derived clastic sediments (Rockcliffe Formation) was deposited during Chazyan time (Williams 1991). Following deposition of these sediments, erosion resumed in eastern Ontario until the Blackriverian (Chart D, Map 2582, map case).

The continental shelf upon which Sequence 1 deposits accumulated was progressively deformed by the onset of convergent plate tectonism beginning in the Middle Ordovician. This event, termed the Taconic Orogeny (Rodgers 1971), produced an orogenic belt along the eastern margin of the craton. The Appalachian Basin formed as a marginal trough west of the orogenic belt (Figure 3, Map 2582, map case; Milici and de Witt 1988).

The deformation and destruction of the continental shelf, initiated during Whiterockian to Chazyan time in the Eastern St. Lawrence Platform (Quebec), led to the emplacement of Taconic allochthons by the close of the Chazyan (James and Stevens 1982; Williams and Smyth 1983). As the orogeny progressed, the earlier southeastward transport of craton-derived clastic sediments from the shield (during Sequence 1 and early Sequence 2) gave way to the northwestward transport of orogen-derived clastic sediments into southern Ontario from the Taconic Highlands in the southeast (Sanford, in press). In the Western St. Lawrence Platform (southern Ontario), deposition of minor amounts of orogen-derived clastic sediments began in late Middle Ordovician (Trentonian) time (Figure 3, Map 2582, map case). The main pulse of orogen-derived clastic sediments from the Taconic Orogeny was deposited in southern Ontario during Sequence 3.

During this period of deformation, starting in the Middle Ordovician (Blackriverian), southern Ontario was again inundated by an inland sea, which persisted for much of the remainder of the Ordovician. The stratigraphy of Paleozoic outliers in Ontario (Hume 1925; Kay 1942; Bolton and Copeland 1972; Barnes et al. 1981; Russell 1984) and adjacent Quebec (Clark 1972; Walters et al. 1976; Harland et al. 1985) indicates progressive overlap of Precambrian Shield, Sequence 1 and older Sequence 2 rocks by Blackriverian, Trentonian and ultimately Edenian sediments, during the Sequence 2 transgression (Figure 3, Map 2582, map case). This inundation produced a succession of shallow-water marine carbonate rocks and minor craton-derived and orogen-derived clastic rocks collectively known as the Black River and Trenton groups in southwestern Ontario, the upper part of the Basal Group and the Simcoe Group in south-central Ontario, the Ottawa Group in eastern Ontario, and the lower part of the Liskeard Group in the Lake Timiskaming outlier (Figure 20.11; Wilson 1946; Sanford 1961; Liberty 1969; Russell 1984; Williams 1991).

The widespread inundation of the southern Canadian Shield during the Trentonian, Edenian and Maysvillian extended, presumably via the Williston Basin in southern Manitoba, into northern Ontario, marking the inception of the Hudson Bay and Moose River basins (Figure 3 and Chart D, Map 2582, map case). The oldest sedimentary rocks in these basins (the Bad Cache Rapids Group and Boas River Formation) were deposited during the final phase of this depositional sequence (Sanford 1987).

Figure 20.11 gives the stratigraphic nomenclature used in Ontario for the units of Sequence 2. Lithologic details of these units are shown in Table 20.2. The distribution of the units comprising Sequence 2 are shown on Maps 2541 to 2544 (map case); the distribution of the sequence as a whole is shown on Maps 2575 to 2578 (map case). The previous maximum distribution of this sequence and of its major lithologic components is shown on Figure 3, Map 2582 (map case). In the following section, details of the stratigraphy of this sequence are presented, subdivided on a geographic basis.

# Geology

## EASTERN AND SOUTH-CENTRAL ONTARIO

The oldest deposits of Sequence 2 in Ontario are Chazyan stage rocks of the Rockcliffe Formation (also called the Chazy Group; *see* Figure 20.11; Chart D, Map 2582, map case; Wilson 1946; Williams 1991), which are confined to the Ottawa Embayment (Map 2544; Figure 3, Map 2582, map case). These predominantly craton-derived clastic rocks disconformably overlie rocks of Sequence 1 and were derived from the uplifted adjacent Frontenac and Laurentian arches (Sanford, in press).

Rockcliffe Formation strata outcrop sporadically in the central and northern parts of eastern Ontario (*see* Figure 20.8) and are subdivided into an upper (St. Martin) member and a lower (unnamed) member (Williams 1991). Both members contain interbedded quartzitic sandstone and shale, with the upper member also including limestone and silty dolostone interbeds (*see* Table 20.2; Williams 1991). Periodic exposure during deposition is indicated by desiccation cracks. A supratidal to subtidal shelf depositional environment is interpreted for this unit (Williams 1991), with deeper water conditions to the east indicated by trace fossil assemblages (Hofmann 1979). The Rockcliffe Formation is disconformably overlain by sandstones and dolostones of the Shadow Lake Formation.

The remainder of the Sequence 2 interval is represented in southern Ontario by a complex succession of shallow marine carbonate rocks and minor clastic rocks which were deposited by a marine transgression subsequent to the hiatus following Rockcliffe Formation deposition. Lateral facies changes in this succession have resulted in a plethora of

| | | | Northern Ontario | | | Southern Ontario | | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Series | Stage | | Hudson Bay and Moose River basins | Lake Timiskaming outlier | | Southwestern Ontario (subsurface) | | South-Central Ontario | | Eastern Ontario | |
| Upper Ordovician | Maysvillian | | Boas River Fm. | | | Trenton Group | Cobourg Fm. | Simcoe Group | Collingwood Mb. | Ottawa Group | Eastview Mb. |
| | Edenian | | Bad Cache Rapids Group | Liskeard Group | Farr Fm. | | | | Lindsay Fm. | | Lindsay Fm. |
| Middle Ordovician | Trentonian | Shermanian | ? | | Bucke Fm. | | Sherman Fall Fm. | | Verulam Fm. | | Verulam Fm. |
| | | Kirkfieldian | | | | | Kirkfield Fm. | | Bobcaygeon Fm. | | Bobcaygeon Fm. |
| | | Rocklandian | | | | Black River Group | Coboconk Fm. | | | | |
| | Blackriverian | | | | ? Guigues Fm. | | Gull River Fm. | | Gull River Fm. | | Gull River Fm. |
| | | | | | | | Shadow Lake Fm. | Basal Group | Shadow Lake Fm. | | Shadow Lake Fm. |
| | Chazyan | | | ? | | | | | | | Rockcliffe Fm. |

Figure 20.11. Stratigraphic units of Sequence 2 (? indicates uncertain age).

**Table 20.2.** Middle Ordovician (Whiterockian) to Late Ordovician (Maysvillian) units in Ontario—Depositional Sequence 2 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Southern Ontario** | | | | | | |
| Lindsay Formation (Simcoe, Ottawa and Trenton groups) | Eastern, south-central and southwestern (subsurface) Ontario | Trentonian to Maysvillian | Shallow shelf to shoal environment (lower member) grading into deep shelf (upper member) | max. 67 m | Upper member: Collingwood Mb. (southwest and central Ontario); up to 10 m of interbedded, black, organic-rich limestone and highly calcareous black shale; fossiliferous (significant trilobite, graptolite and conularid fauna).<br>Eastview Mb. (eastern Ontario) is similar in lithology and thickness to Collingwood Mb.<br>Lower member (unnamed): argillaceous, nodular, fine- to coarse-grained limestone; very fossiliferous (significant trilobite fauna); intensely bioturbated, giving nodular appearance; normal grading, cross-lamination.<br>Upper contact of Lindsay Fm. with Blue Mountain Fm. (Sequence 3) in central and southwestern Ontario is commonly disconformable and locally marked by presence of a phosphate bed; lower contact with Verulam Fm. is gradational. | Ludvigsen 1979b;<br>Coniglio et al. 1990;<br>Russell and Telford 1983;<br>Brookfield and Brett 1988;<br>Liberty 1969;<br>Williams 1991;<br>Harris 1985;<br>Churcher et al. 1991 |
| Verulam Formation (Simcoe, Ottawa and Trenton groups) | Eastern, south-central and southwestern (subsurface) Ontario | Trentonian | Open, normal marine shelf or ramp; shallow shoal environment in upper part of unit | 32 to 65 m | Interbedded, micritic to coarse-grained, fossiliferous limestone with inter-beds of calcareous shale; light to dark grey, brown-grey; thin to medium bedded; bioturbation; intraclasts; ripple and trough cross-lamination, channels and graded beds. Lower contact gradational with Bobcaygeon Fm. | Liberty 1969;<br>Williams 1991;<br>Coniglio et al. 1990;<br>Brookfield and Brett 1988 |
| Bobcaygeon Formation (Simcoe, Ottawa and Trenton groups) | Eastern, south-central and southwestern (subsurface) Ontario | Blackriveran to Trentonian | Shallow subtidal, well-agitated environment | 7 to 87 m | Dark to light grey, brown to blue-grey, interbedded micritic to coarse-grained, fossiliferous limestone; variable argillaceous content; planar to undulating shaly partings; micritic beds are subordinate; "bird's-eye" texture in fine-grained beds; planar ripple and trough cross-lamination; graded beds; bioturbation and stylolites in coarse-grained beds. Lower contact sharp to gradational with Gull River Fm. | Williams 1991;<br>Liberty 1969;<br>Barnes 1967;<br>Coniglio et al. 1990;<br>Brookfield and Brett 1988 |

Table 20.2. Middle Ordovician (Whiterockian) to Late Ordovician (Maysvillian) units in Ontario—Depositional Sequence 2 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Gull River Formation (Black River, Simcoe and Ottawa groups) | Eastern, south-central and southwestern (subsurface) Ontario | Blackriveran | Hypersaline supratidal to intertidal flat, grading upward into restricted to more open marine, subtidal lagoon | 7.5 to 136 m | Upper member: grey to brownish grey, fine-grained to micritic limestone; "bird's-eye" structures.<br>Lower member: light brown-grey to blue-grey interbedded limestone and silty dolostone (grading to siltstone), with shaly partings; thin to thick bedded; minor quartz sandstone interbeds.<br>Desiccation cracks, planar to wavy laminations and ripple marks; abundant fossils; sulphate nodules and molds; burrows; stylolites.<br>Lower contact is conformable and gradational with Shadow Lake Fm. and is defined by lowest occurrence of pure carbonate bed. | Okulitch 1939;<br>Liberty 1969;<br>Williams 1991;<br>Barnes 1967;<br>Mukherji 1969;<br>Coniglio et al. 1990 |
| Shadow Lake Formation (Black River, Basal and Ottawa groups) | Eastern, south-central and southwestern (subsurface) Ontario | Blackriveran | Nearshore (eastern Ontario); alluvial and/or deltaic deposits reworked in nearshore to supratidal environment (south-central Ontario) | generally 2 to 3 m; 15 m max. | Central and southwestern Ontario: interbedded red to green, silty, dolomitic or calcareous sandstones and terrigenous mudstones; occasionally silty and/or sandy dolostone interbeds becoming more common upward.<br>Eastern Ontario: calcitic to noncalcitic, silty to sandy, light-grey to green-grey, fine to microcrystalline dolostone; interbeds of fine-grained calcareous sandstone.<br>Planar and ripple cross-lamination, desiccation cracks, burrows; fossils very rare.<br>Lower contact is disconformable with the Rockcliffe Fm. (eastern Ontario) and Cambrian rocks (central and southwestern Ontario) and unconformable with the Precambrian. | Liberty 1969;<br>Williams 1991;<br>Trevail 1990;<br>Sanford 1961;<br>Barnes 1967;<br>Mukherji 1969;<br>Coniglio et al. 1990 |
| Rockcliffe Formation (a.k.a. Chazy Group) | Eastern Ontario (Ottawa Embayment) | Chazyan | Supratidal to subtidal shelf | up to 125 m | Lower member (unnamed): interbedded, light grey to green, fine-grained, calcareous to noncalcareous quartz sandstone and dark grey to maroon shales; basal quartz-pebble conglomerate locally present.<br>Upper (St. Martin) member: same rock type as lower but also includes interbeds of shaly bioclastic limestone and silty dolostone.<br>Desiccation cracks, cross-bedding, ripple marks, flute casts and burrows.<br>Lower contact is disconformable with units of Sequence 1 (*see* Table 20.1). | Williams 1991;<br>Wilson 1946;<br>Hofmann 1979 |

Table 20.2. Middle Ordovician (Whiterockian) to Late Ordovician (Maysvillian) units in Ontario—Depositional Sequence 2 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Northern Ontario** | | | | | | |
| Boas River Formation | Hudson Bay Basin | Maysvillian? | Slow deposition within stratified basin | up to at least 17 m | Upper unit: brownish black to dark brown, fissile, fine-laminated, organic-rich limestone.<br>Lower unit: light to medium brown, poorly laminated, organic-rich limestone interbedded with organic-poor, light yellow-brown limestone.<br>Graptolites, trilobites and conodonts.<br>Upper and lower contacts with overlying Churchill River Gp. (Sequence 3, Table 20.3) and underlying Bad Cache Rapids Gp., are disconformable(?) and conformable, respectively. | Macauley 1986;<br>Macauley et al. 1990;<br>Sanford and Grant 1990;<br>Heywood and Sanford 1976 |
| Bad Cache Rapids Group | Hudson Bay and Moose River basins | Trentonian? to Edenian | Intertidal to subtidal environment | 50 to 200 m | Basal calcareous sandstone and siltstone grading up into poorly fossiliferous and nodular, yellow-grey, fine-grained, cherty limestone.<br>Lower contact unconformable with Precambrian. | Sanford 1987;<br>Sanford et al. 1968;<br>Norris 1986 |
| Farr Formation (Liskeard Group) | Lake Timiskaming outlier | Trentonian to Maysvillian | Undetermined | up to 41 m | Medium- to thick-bedded, fine- to medium-grained, fossiliferous calcarenites, becoming finer grained and dolomitic at top of unit; undulating, shaly stringers; intensely burrowed giving unit a mottled appearance.<br>Lower contact with the Bucke Fm. is gradational.<br>Upper contact with Dawson Point Fm. (Sequence 3) is disconformable. | Russell 1984;<br>Copeland 1965;<br>Munro 1975 |
| Bucke Formation (Liskeard Group) | Lake Timiskaming outlier | Trentonian | Undetermined | up to 20 m | Soft, fissile, green and blue to grey shale alternating with nodular, fine-grained to micritic, shaly limestone. Fossiliferous in upper part. Lower contact with Guigues Fm. is gradational(?). | Russell 1984 |
| Guigues Formation (Liskeard Group) | Lake Timiskaming outlier | Blackriveran to Trentonian | Undetermined | up to 51 m | Fine- to medium-grained sandstone grading upward into red, green and grey noncalcareous siltstone. Locally rounded quartz-syenite boulders (up to 2 m in diameter) in dolomitic matrix, poorly sorted red quartz-pebble conglomerate and coarse-grained sandstone occur.<br>Lower contact unconformable with Precambrian rocks. | Russell 1984;<br>Hume 1925;<br>Thomson 1965 |

stratigraphic nomenclature for the Middle and lower Upper Ordovician strata in Ontario (Wilson 1946; Sanford 1961; Beards 1967; Liberty 1969; Winder and Sanford 1972; Russell and Telford 1983; Williams 1991). The variety of nomenclature reflects differing emphasis on various criteria such as biostratigraphy, lithostratigraphy, geophysical properties and weathering characteristics. This nomenclature has been most recently reviewed by Noor (1989), Coniglio et al. (1990) and Williams (1991). The nomenclature currently used by the OGS in the outcrop belt is shown in Figure 20.11 and is based on that of Liberty (1969) as modified by Russell and Telford (1983), Williams and Telford (1986) and Williams (1991).

Basal sediments of the Middle Ordovician transgression are represented by the Shadow Lake Formation, a distinctly diachronous unit, consisting of craton-derived clastic rocks and impure carbonate rocks, deposited on the irregular topography of eroded older strata (Liberty 1969). In south-central Ontario, the Shadow Lake Formation has been assigned to the Basal Group (*see* Figure 20.11; Liberty 1969). This is because of its lithologic similarity to underlying Cambrian strata and despite its disconformable lower contact (Trevail 1990) with Cambrian units, and its conformable upper contact with the Gull River Formation of the Simcoe Group (Liberty 1969). In eastern Ontario, the Shadow Lake Formation is included with overlying Middle to Upper Ordovician units of the Ottawa Group (*see* Figure 20.11).

In eastern Ontario, the Shadow Lake Formation disconformably overlies the Rockcliffe Formation (*see* Figure 20.11; Williams 1991). In south-central and southwestern Ontario, the Shadow Lake Formation either unconformably overlies the Precambrian basement or disconformably overlies the predominantly clastic Cambrian units of Sequence 1, where present. This unit commonly reflects a local provenance: where it overlies gneissic Precambrian terrain, it has a higher clastic component (i.e., feldspathic and conglomeratic sandstone); where it overlies a carbonate substrate (i.e., Cambrian limestone or Precambrian marble), the formation is more calcareous (Sanford 1961; Liberty 1969; Williams 1991).

Typically, the Shadow Lake Formation consists of a few metres of dolomitic and sandy shale (*see* Table 20.2). However, its thickness and grain size may vary considerably depending on local pre-existing topography (Sanford 1961). Depositional environments proposed for the Shadow Lake Formation include nearshore (eastern Ontario) and supratidal (south-central Ontario) environments (Barnes 1967; Mukherji 1969). Its contact with the overlying Gull River Formation is commonly gradational and is placed at the first significant carbonate bed (Liberty 1969; Williams 1991).

The overlying succession of limestones and shales of the Gull River, Bobcaygeon, Verulam and Lindsay formations represent relatively continuous deposition on a generally deepening shelf (Brookfield and Brett 1988; Coniglio et al. 1990). These units comprise the Simcoe Group in south-central Ontario (Liberty 1969) and, with the Shadow Lake Formation, form the Ottawa Group in eastern Ontario (*see* Figure 20.11). The rock types in this succession range from coarse-grained bioclastic carbonates (grainstones and calcarenites) to carbonate mudstones (microcrystalline or lithographic), with subordinate calcareous and non-calcareous shales. Although each unit has general identifying characteristics, precise formational identification in the field is difficult because of repetitive facies and rapid lateral and vertical facies changes.

The Gull River Formation (Okulitch 1939; Liberty 1969; Williams 1991) is typically a limestone which is locally argillaceous, silty or dolomitic. This unit has been subdivided into 2 members (Williams 1991), the lowermost being more silty and dolomitic (*see* Table 20.2). Strata previously recognized as the upper member of the Gull River Formation (e.g., Liberty 1969; Carson 1982) have been reassigned to the overlying Bobcaygeon Formation (Derry, Michener, Booth and Wahl and OGS 1989a; Williams 1991).

Common to both the upper member of the Gull River Formation and the overlying Bobcaygeon Formation (lower member) are "birds-eye" structures consisting of small, white, horizontally elongated calcite lenses scattered throughout the microcrystalline limestone beds (Williams 1991). Depositional features include desiccation cracks, planar to wavy laminations, burrows and intraformational conglomerates. Some Gull River Formation strata are richly fossiliferous, containing brachiopods, bryozoa, corals, ostracods, gastropods, cephalopods, trilobites and stromatolites (Liberty 1969; Williams 1991). The shelly fauna confirm a Blackriverian age for this formation (Barnes et al. 1981).

The Gull River Formation represents deposition within a supratidal to intertidal flat environment with coarser-grained beds representing storm deposition (Barnes 1967; Mukherji 1969; Coniglio et al. 1990). Minor quartzitic sandstone interbeds indicate that the supply of clastic detritus from local Precambrian topographic highs was available at least until deposition of the upper part of the formation (Williams 1991).

Excellent exposures of the Gull River Formation and the underlying Shadow Lake Formation can be seen at the abandoned Marmoraton Iron Mine (Figure 20.12) near Marmora in south-central Ontario (Derry, Michener, Booth and Wahl and OGS 1989b; Noor 1989; Coniglio et al. 1990).

The Bobcaygeon Formation, as defined in eastern and east-central Ontario by Williams (1991), consists mainly of fossiliferous limestone, with variable argillaceous content. It is subdivided into 3 members based on grain size (Liberty 1969) or shale content (Williams 1991). The Bobcaygeon Formation is generally more fossiliferous than the Gull River Formation, with abundant brachiopods, bryozoa, corals, gastropods, large cephalopods, stromatoporoids, trilobites, ostracods and conodonts of Blackriverian age (Barnes 1967; Barnes et al. 1981; Williams 1991). The fauna, grain size and sedimentary features (*see* Table 20.2) suggest a shallow subtidal depositional environment (Barnes 1967; Coniglio et al. 1990).

The Verulam Formation consists of limestone interbedded with calcareous shale (Figure 20.13; *see* Table 20.2;

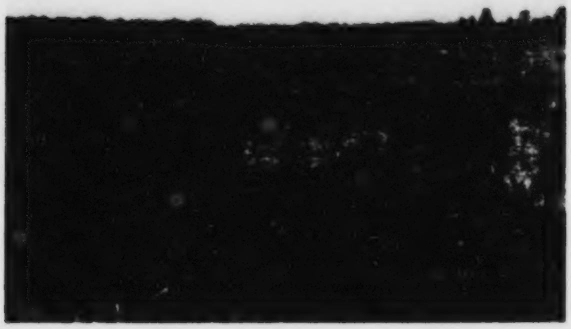

**Figure 20.12.** Middle Ordovician limestones and minor sandstones and shales (light grey upper 3 benches in photograph) overlying Precambrian (Grenville) metasedimentary rocks at the Marmoraton Iron Mine, near Marmora. Up to 55 m of these Sequence 2 units (the Shadow Lake, Gull River and Bobcaygeon formations) are exposed in this section. (Photo: DA90–22–25)

**Figure 20.13.** Interbedded limestone and shale of the Verulam Formation at the old Bath Quarry. (Photo: DA90–44–34)

Liberty 1969; Williams 1991). The shale content results in generally poor quality exposures of this unit, except where capped by limestones of the overlying Lindsay Formation. The lower contact of the Verulam Formation with the Bobcaygeon Formation is gradational, identified by an upward increase in shale content and a decrease in maximum bed thickness (Williams 1991). The Verulam Formation is also generally more fossiliferous than the Bobcaygeon Formation strata, containing abundant brachiopods, bryozoa, gastropods and crinoid debris (Liberty 1969; Williams 1991).

The depositional environment of the Verulam Formation was an open normal marine shelf or ramp (Coniglio et al. 1990). The limestone beds tend to increase in grain size upward, suggesting shallowing upward to shoal conditions at the top of the formation (Coniglio et al. 1990). The shale content of this unit may have been derived from erosion of Taconic allochthons to the southeast.

The Lindsay Formation consists of 2 members, a lower unnamed, nodular limestone member, and an upper organic-rich, calcareous shale member (*see* Figure 20.11; *see also* Table 20.2). The upper member is known in southwestern and south-central Ontario as the Collingwood Member and in eastern Ontario as the Eastview Member (Russell and Telford 1983; Williams 1991).

The origin of the characteristic nodular texture of the lower member of the Lindsay Formation is unknown, although it is most likely related to a combination of bioturbation, diagenesis and compaction (Coniglio et al. 1990). The Lindsay Formation is richly fossiliferous and includes a significant trilobite fauna (Ludvigsen 1979b). The faunal content and bioturbated nature of the lower member indicates deposition in an open, shallow to deep shelf setting (Brookfield and Brett 1988; Coniglio et al. 1990).

The Collingwood Member is a relatively thin unit which contains a significant well-preserved trilobite, graptolite and conularid fauna (Figure 20.14). Most notable are the trilobites *Pseudogygites canadensis* and *Triarthrus* (Ludvigsen 1979b; Tuffnell and Ludvigsen 1984). Good fossil collecting areas occur along the Georgian Bay shoreline northwest of Craigleith (Verma 1979).

Collingwood Member strata, previously assigned to the Craigleith Member of the Whitby Formation (Liberty 1969), were reassigned to the Lindsay Formation by Russell and Telford (1983), based on the gradational contact of these strata with underlying Lindsay Formation strata and on their calcareous nature. Russell and Telford (1983) renamed the remaining Whitby Formation strata, the Blue Mountain Formation (*see* Figure 20.15; Sequence 3). In addition, Russell and Telford (1983) clarified that strata in the subsurface of southwestern Ontario, which was assigned to the Collingwood Formation (of Sanford 1961), should be reassigned to the Blue Mountain Formation.

In southern Ontario, the Collingwood Member is geographically restricted to a zone extending from the Toronto–Oshawa area, east to Lake Huron, and north to the Bruce Peninsula and Manitoulin Island (Russell and Telford 1983; Churcher et al. 1991). A number of subsurface outliers of this unit occur to the southwest, suggesting a more widespread initial distribution (Churcher et al. 1991). Post-depositional erosion of the Collingwood Member strata is supported by the presence of an upper erosional contact in some drill cores (Russell and Telford 1983) and by the occurrence of a phosphatic lag bed between the Lindsay and Blue Mountain formations where the Collingwood Member is absent (Churcher et al. 1991).

The Collingwood Member is commonly referred to as an "oil shale" because of its high organic carbon content (up to 13%) and the relative thermal immaturity of this organic matter (Macauley et al. 1985, 1990). This unit is 1 of 3 southern Ontario oil shales investigated by the Ontario Geological Survey in the early 1980s as part of the Oil Shale Assessment Project (Churcher et al. 1991; Barker, Dickhout et al. 1983; Barker, Russell et al. 1983; Johnson 1983; Johnson et al. 1983a, 1983b; *see also* Economic Geology and Sequences 7 and 8). Detailed organic geochemical analyses (Barker, Dickhout et al. 1983; Barker, Russell et al. 1983) indicate that generally more than half of the organic matter in the Collingwood Member is kerogen. The

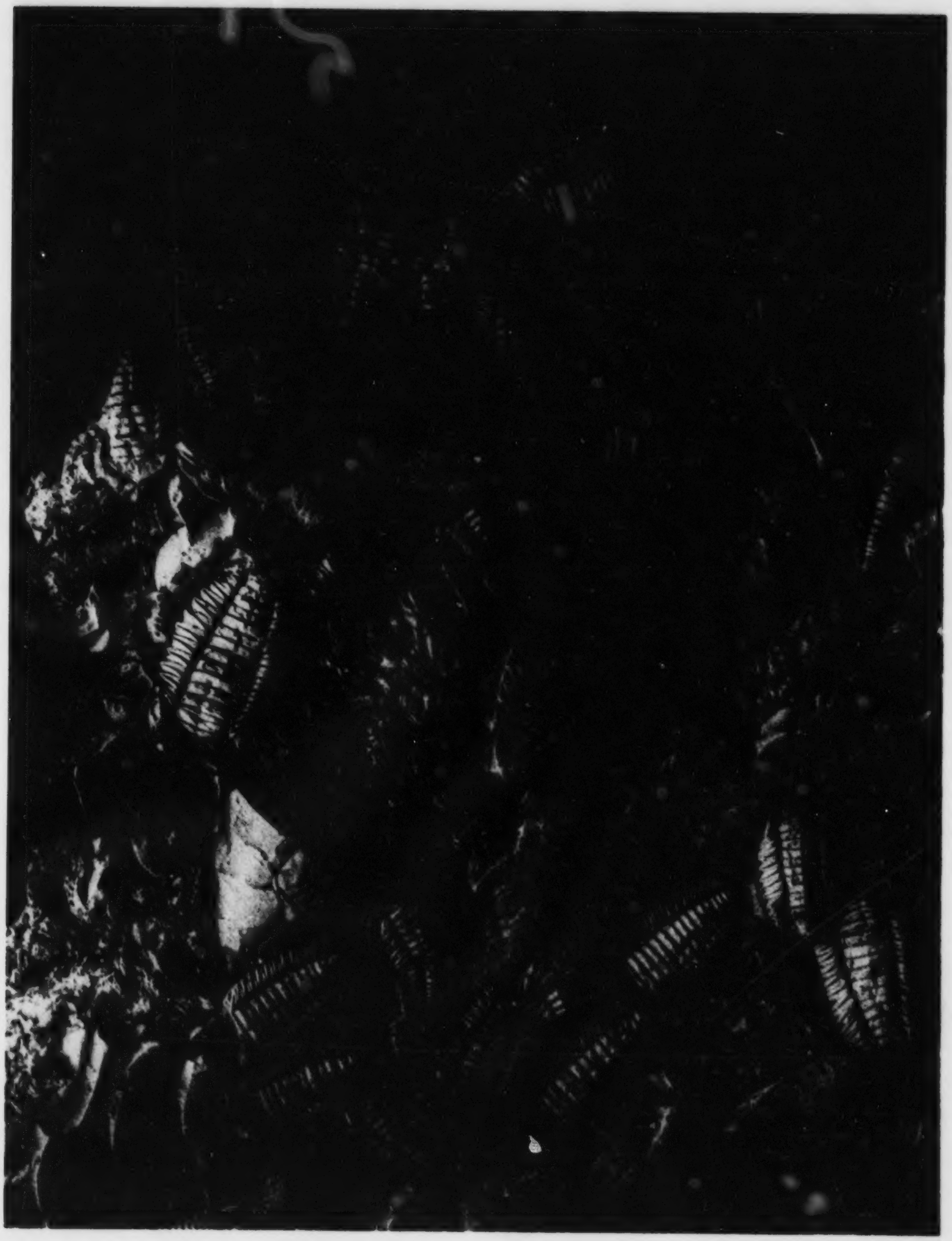

**Figure 20.14.** Abundant trilobites (*Triarthrus eatoni*) in a slab of the Collingwood Member, Lindsay Formation, from Craigleith (ROM 35375). This specimen is unusual in that whole trilobites, ranging in size from 1.35 to 2 cm, have been preserved (*from* Tuffnell and Ludvigsen 1984).

remaining bituminous content is evidenced by the petroliferous odour given off when fresh samples are first broken.

In eastern Ontario, organic-rich argillaceous limestones at the top of the Lindsay Formation and of approximately the same age as the Collingwood Member, are referred to as the Eastview Member of the Lindsay Formation (*see* Figure 20.11; Williams and Telford 1986), formerly the Eastview Formation of Liberty (1963). Eastview Member strata are similar in rock type and thickness to the Collingwood Member and clearly represent the same depositional environment.

The lack of shallow water depositional indicators (e.g., wave ripples or grainstone beds) suggest that the Collingwood and Eastview members were deposited in a deep shelf setting (Coniglio et al. 1990). It is generally accepted that organic matter in oil shales is syngenetic and that the anomalous quantities of organic matter are due to high biologic activity in surface waters (Pederson and Calvert 1990) and/or deposition under anoxic conditions (Demaison and Moore 1980; *see also* Sequence 8). The presence of approximately coeval (lower Upper Ordovician), organic- rich units in widely scattered locations (also Boas River Formation in the Hudson Platform), suggests that the mechanism responsible for these deposits operated on a cratonic scale.

## SOUTHWESTERN ONTARIO

In the subsurface of southwestern Ontario, a biostratigraphically based nomenclature is used for the interval equivalent to the Ottawa and Simcoe groups (*see* Figure 20.11). These strata are subdivided into the Trenton and Black River groups (Sanford 1961; Liberty 1969) and their approximate correlative relationships with the units of south-central and eastern Ontario are illustrated in Figure 20.11. Although defined biostratigraphically, these units have been informally assigned a lithostratigraphic character based on geophysical logs and well cuttings (Sanford 1961; Beards 1967; Brigham 1971; Bailey and Cochrane 1984b). The Trenton–Black River contact is a convenient subsurface marker, normally identified by a "spike" on the gamma-ray log. This shaly marker bed falls approximately in the middle of the Bobcaygeon Formation and is not coincident with the Blackriverian–Trentonian boundary (*see* Figure 20.11).

The thickness of Middle to lower Upper Ordovician (i.e., Sequence 2, excluding the Rockcliffe Formation) interval in the subsurface of southwestern Ontario ranges up to approximately 300 m (Johnson et al. 1985; Winder and Sanford 1972).

## NORTHERN ONTARIO

The Middle to lower Upper Ordovician trangression extended over much of the Canadian Shield (Map 2542; Figure 3 and Chart D, Map 2582, map case), as indicated by the presence of the Liskeard Group in the Lake Timiskaming outlier (Russell 1984) and Middle Ordovician strata in the Lake Nipissing outlier (Barnes et al. 1981). Time equivalent strata were also deposited on the Hudson Platform (Bad Cache Rapids Group and Boas River Formation), but there is no evidence to indicate a direct seaway linkage between these 2 areas.

The basal Guigues Formation of the Liskeard Group consists of poorly sorted conglomerates and sandstones, locally grading into siltstones (Russell 1984). The unit is analogous to the Shadow Lake Formation of southern Ontario and reflects initial transgression across the Precambrian Shield.

The Guigues Formation is overlain by the shales and nodular limestones of the Bucke Formation, which are in turn overlain by calcarenites of the Farr Formation (Russell 1984). All of these units are included in the Liskeard Group (*see* Figure 20.11). The Farr Formation is richly fossiliferous in corals, brachiopods, gastropods and the unusual fossil *Receptaculites* (Hume 1925; Russell 1984). Conodonts indicate the Farr Formation to be of Trentonian to early Maysvillian age (Copeland 1965; Munro 1975).

Middle Ordovician strata are also preserved in an outlier on an island in Lake Nipissing, near North Bay (Map 2544, map case). The mainly carbonate succession of this outlier is biostratigraphically correlated with the Guigues and Bucke formations of the Lake Timiskaming outlier and with the Bobcaygeon Formation of southern Ontario (Barnes et al. 1981).

In the Hudson Platform, initial Paleozoic deposition is represented by the Middle to Upper Ordovician Bad Cache Rapids Group (Sanford 1987). This unit consists of a basal calcareous sandstone and siltstone grading up into cherty limestone and is interpreted to have been deposited in intertidal to subtidal environments (Sanford et al. 1968; Cumming 1975; Sanford 1987).

Overlying the Bad Cache Rapids Group, in the Hudson Platform, is an organic-rich limestone unit, the Boas River Formation, which may correlate with the Collingwood and Eastview members (Lindsay Formation) of southern Ontario (Heywood and Sanford 1976; Macauley 1986; Macauley et al. 1990). This unit is exposed at Boas River, Southampton Island in the Northwest Territories and on Akapatok Island in Ungava Bay, Quebec (Heywood and Sanford 1976). Recent drilling in the Hudson Bay Basin has delineated this interval in the subsurface, and led Sanford and Grant (1990) to formally introduce the term Boas River Formation into Ontario. This unit was not differentiated on Maps 2542 to 2544, where it should be located in the vicinity of the plotted contact between the Bad Cache Rapids and Churchill River groups.

# ECONOMIC GEOLOGY

Conventional oil and gas resources of the Middle to Upper Ordovician Sequence 2 rocks of southwestern Ontario are mainly found in association with dolomitized fractures, faults and bedding planes. Dolomitizing fluids migrating through these conduits increased the porosity and permeability of the surrounding strata, forming vertical and horizontal "sheet-like traps" (Brigham 1971; Bailey and Cochrane 1984b; Middleton et al. 1990).

Oil and gas production from Ordovician strata in southwestern Ontario has shown a significant increase since the mid–1980s, due, in part, to the application of tectonic fracture-framework exploration models (e.g., Sanford et al. 1985; *see* Summary and Discussion). Total oil production for 1989 was 177 188 m³, accounting for 72.7% of the total Ontario production (Carter 1990). Cumulative oil production (through to the end of 1989) from Ordovician strata amounted to 662 000 m³, or 6.7% of the total cumulative Ontario oil production. Total gas production for 1989 was 27 159 thousand m³, or 5.5% of the total production in the province. Cumulative gas production to the end of 1989 amounted to 463 100 thousand m³, or about 1.6% of the total cumulative Ontario gas production (Carter 1990).

The oil shales of the Collingwood Member (Lindsay Formation) represent potential unconventional hydrocarbon resources. In 1859, a fledgling oil shale industry near Craigleith, Ontario, briefly produced shale oil from strata of the Collingwood Member (Macauley 1984). The retorting process was costly, crude and dangerous, with the works being destroyed by fire on several occasions. The oil shale works were rendered uneconomic by the early 1860s with the commercial production of cheaper conventional crude oil from North America's first oil well completed in Devonian strata near Oil Springs, Ontario (*see* Economic Geology of Sequence 7).

Recent studies by the OGS and others have determined that the higher grade intervals from the Collingwood Member are capable of yielding in the order of 14 to 30 litre per tonne (l/t) of crude shale oil using normal retorting techniques (Barker, Dickhout et al. 1983; Barker, Russell et al. 1983; Barker 1985; Macauley et al. 1990; Churcher et al. 1991). The application of more advanced processes such as "HYTORT" (i.e., retorting in a hydrogen-rich environment) could approximately double this yield (Barker, Dickhout et al. 1983).

The thin (typically less than 3 m) and interbedded nature of the organic-rich beds limit this unit's resource potential. However, given high oil prices, similar to those of the 1970s and early 1980s, the economic potential of the Collingwood Member and the Upper Devonian Kettle Point oil shale (*see* Sequence 8) could become more significant.

In eastern Ontario, the Eastview Member (Lindsay Formation) has a low potential as a source of shale oil, because its organic matter is thermally overmature (Barker, Dickhout et al. 1983). The anomalous maturation of the Eastview Member organic matter may have been caused by higher heat flow related to passage of this part of the North American plate over a mantle hotspot during Cretaceous to early Tertiary times (Legall et al. 1981; *see also* Sequence 10 and Summary and Discussion).

In the Hudson Platform, the Boas River Formation represents a potential hydrocarbon source rock. Macauley et al. (1990) stated that if this unit were buried to a sufficient depth (i.e., attained thermal maturity), it would constitute an excellent source rock.

With the exception of the Shadow Lake Formation (which has limited use as general fill), all the formations present in the Simcoe and Ottawa groups are important stone resources to the province (Bezys and Johnson 1988; Derry, Michener, Booth and Wahl and OGS 1989a, 1989b, 1989c). Gull River, Bobcaygeon and Lindsay formation strata are all used for general aggregate purposes. The Verulam Formation is used mainly in cement manufacturing, due to its high shale content. Lindsay Formation strata are also a significant source of raw material for cement manufacture.

From an aggregate quality viewpoint, alkali-carbonate reactivity is a significant problem associated with certain horizons in Sequence 2 (Dolar-Mantuani 1975, 1983; Rogers 1985, 1986). This phenomenon causes expansion (up to several volume%) and degradation of concrete due to the reaction of dolomitic limestone aggregate of a relatively narrow range of chemical composition with alkalis in cement. The expansion causes cracking and buckling, and commonly results in the offset of structures (e.g., curbs and drains). Aggregates susceptible to alkali-carbonate reactivity have been traced to specific horizons within the upper Gull River and lower Bobcaygeon formations in the Lake Simcoe, Kingston and Ottawa areas (Rogers 1985). The precise geographic distribution of the alkali reactive strata is still poorly known.

The Gull River Formation has been a major historical source of building stone, particularly in the Ottawa–Cornwall and Kingston areas (Derry, Michener, Booth and Wahl and OGS 1989b). A large number of stone structures have been built with this material.

In the Timiskaming area, the Farr Formation is a significant source of crushed stone and is utilized as a chemical flux stone (Grant and Owsiacki 1987).

# DEPOSITIONAL SEQUENCE 3: LATE ORDOVICIAN (MAYSVILLIAN TO GAMACHIAN)

## Overview

Sequence 3 is characterized by orogen-derived clastic rocks deposited in marine to nonmarine settings. The lower boundary of Sequence 3 is placed at the locally disconformable contact between the dominantly carbonate succession of Sequence 2 and the clastic succession of Sequence 3. The upper boundary of Sequence 3 is an erosional unconformity of Gamachian age, equivalent to the Tippecanoe I–II subsequence boundary of Sloss (1988).

The Taconic Orogeny of eastern North America, coupled with a possible glacio-eustatic sea level drop related to a continental glaciation on the North African craton (Dennison 1976), profoundly affected the character and distribution of sediments in the Appalachian and Michigan basins. Continued northwestward migration of the Appalachian structural front in the Late Ordovician (Maysvillian to Gamachian stages) provided a major source of orogen-derived clastic detritus to the eastern North American craton. This clastic detritus was deposited progressively farther westward during this period (Sanford, in press). The Trentonian to Maysvillian shale, flysch and carbonate units deposited along the outer margins of the St. Lawrence Platform (in New York State and Pennsylvania) in the previous sequence, were eroded and redeposited across southern Ontario and at least as far north as Lake Timiskaming (Figure 4 and Chart D, Map 2582, map case; Russell 1984). The resulting strata of Sequence 3 consist mainly of orogen-derived marine shales with a lesser amount of carbonate rocks.

As the Appalachian structural front reached its most westerly extent, the marginal trough (northwest of the uplift) was segmented into the Quebec and Appalachian basins (Sanford, in press). The marine shales in southern Ontario were succeeded by the nonmarine to marginally marine sediments of the Queenston delta complex, an extensive redbed succession shed from the Taconic Highlands at the culmination of the Taconic Orogeny. Very shallow marine (and possibly nonmarine) sediments were deposited in southern Ontario, whereas more normal marine sediments were deposited to the north and west. These latter marine clastic rocks interfinger laterally with carbonate rocks of the Michigan Basin (Figure 4 and Chart D, Map 2582, map case) exposed on Manitoulin and the other North Channel Islands.

An equivalent succession of marine to marginal marine shales with minor carbonate rocks occurs in the Lake Timiskaming outlier of northern Ontario (Map 2542; Chart D, Map 2582, map case; Russell 1984).

The Maysvillian–Richmondian seaway was continuous with the Williston Basin in southern Manitoba, and may have extended northward across the Severn Arch to connect with the Hudson and Arctic platforms (Figure 4, Map 2582, map case; Sanford 1987). In this area, Sequence 3 consists mainly of limestones and dolostones with halites (Heywood and Sanford 1976; Sanford 1987; Sanford and Grant 1990). Coarse clastic rocks of cratonic origin intertongue with dolostones of intertidal and supratidal origin along the southern margin of the Moose River Basin, suggesting that the adjacent Fraserdale Arch was emergent and tectonically active during the deposition of this sequence (Sanford 1987).

Sequence 3 is upper Maysvillian to Gamachian in age; however, strata representing the youngest Ordovician (Gamachian and possibly part of the Richmondian) are not present in or adjacent to Ontario (Chart D, Map 2582, map case). This gap in the sedimentary record is thought to be related to the Late Ordovician glacio-eustatic sea level drop brought about by continental glaciation on the North African craton (Dennison 1976).

Lithologic details of the stratigraphic units of Sequence 3 are given in Table 20.3. For the distribution of the units, *see* Maps 2541 to 2544 (map case); for the distribution of the sequence as a whole, *see* Maps 2575 to 2578 (map case). A stratigraphic column for this sequence is shown in Figure 20.15. In the following section, details of the stratigraphy of Sequence 3 are presented, and are subdivided on a geographic basis.

## Geology

### SOUTH-CENTRAL AND SOUTHWESTERN ONTARIO

The oldest rocks of Sequence 3 in south-central and southwestern Ontario are the shales of the Blue Mountain Formation (*see* Figure 20.15; Chart D, Map 2582, map case). This unit includes strata of the former Whitby Formation (of Liberty 1969) and of the Collingwood Formation of the subsurface (of Sanford 1961; *see also* Sequence 2; Russell and Telford 1983). The contact with the underlying Collingwood Member (Lindsay Formation) is commonly sharp. Over much of southwestern Ontario, the Collingwood Member was eroded prior to Blue Mountain Formation deposition (Russell and Telford 1983; Churcher et al. 1991). The contact with overlying shales and limestones of the Georgian Bay Formation is gradational and somewhat arbitrarily defined, and is placed at the first appearance of calcareous and/or silty hardbands greater than 5 cm thick (Liberty 1969).

The Blue Mountain Formation typically consists of blue-grey noncalcareous shales, informally termed the Thornbury Member (Russell and Telford 1983). In the Toronto–Pickering area, however, most of the Blue Mountain Formation interval consists of brown to brown-grey shales which have been termed the Rouge River Member (members not shown in Figure 20.15; Russell and Telford 1983). The relationship between these 2 members (or facies) is not well known.

Gradationally overlying the Blue Mountain Formation is the Georgian Bay Formation. The Georgian Bay Formation reflects a gradational change from terrigenous shales with minor siltstones, to shaly and impure carbonates (*see* Table 20.3). Carbonate beds increase in thickness and frequency toward the top of the unit (Figure 20.16); a carbonate cap is locally developed (Liberty and Bolton 1971).

The carbonate content of the Georgian Bay Formation also increases northwestward into the Michigan Basin. On Manitoulin Island, the upper, carbonate-rich part of the Georgian Bay Formation (sometimes referred to as the Kagawong Member) is considered to be laterally equivalent to the Queenston Formation of southern Ontario (Liberty and Bolton 1971; Horst 1978; Barnes et al. 1981; Telford et al. 1981; Byerley and Coniglio 1989).

Most exposures of the Georgian Bay Formation occur below the Niagara Escarpment, northwest of Collingwood (*see* Figure 20.16) and on Manitoulin Island (Map 2544, map case; Liberty and Bolton 1971; Byerley and Coniglio 1989). Exposures through the upper part of the unit on Manitoulin Island, at the Sunova Beach Road and Bridal Veil Falls outcrops, provide some of the best fossil collecting on the island (Telford et al. 1981). The invertebrate fauna are diverse and well preserved; they include solitary and colonial corals, bryozoa, brachiopods and stromatoporoids. The Georgian Bay Formation underlies much of the Metropolitan Toronto area and is commonly exposed (beneath glacial drift) in construction excavations.

Of particular interest in the Georgian Bay Formation is the large number of well-preserved sedimentary structures, such as graded beds, gutter casts, scour-and-fill structures and current- and wave-formed ripple marks (Byerley and Coniglio 1989; Kerr and Eyles 1991). From the study of these features, as well as the biostromes (Horst 1978) and hardground surfaces present, Byerley and Coniglio (1989) have suggested that the formation represents a shallowing upward succession, with storm-controlled depositional patterns.

Many researchers (Foerste 1912, 1924; Parks 1925; Liberty and Bolton 1971; Liberty 1972; Byerley and Coniglio 1989) have attempted to subdivide the strata assigned to this formation into formations or members, based on biostratigraphic or lithostratigraphic criteria. These subdivisions have been largely unsuccessful because of poor outcrop distribution, lateral facies changes and a lack of precise biostratigraphic control. The terms "Meaford" and "Dundas" were at times applied separately or together to describe the strata between the blue-grey shales of the Blue Mountain Formation and the red shales of the Queenston Formation. These terms (e.g., Meaford–Dundas Formation), however, are biostratigraphically based and have now been replaced by the lithostratigraphic term Georgian Bay Formation (Liberty 1969). However, in the subsurface of southwestern Ontario "Meaford–Dundas Formation" is commonly used to describe the strata of both the Blue Mountain and Georgian Bay formations combined (Beards 1967).

The Queenston Formation is the uppermost unit of Sequence 3 in southern Ontario (*see* Figure 20.15; Chart D, Map 2582, map case). This unit is well exposed along the base of the Niagara Escarpment, from the Niagara River to Cabot Head on the Bruce Peninsula, and occurs in the subsurface throughout southwestern Ontario (Sanford 1961; Liberty and Bolton 1971). The Queenston Formation thins northwestwardly, from about 335 m under the eastern part of Lake Erie to about 45 m on the Bruce Peninsula (Sanford 1961).

Overall, the Queenston Formation in Ontario is a shale with interbeds of limestone and calcareous siltstone (Figure 20.17; *see* Table 20.3; Caley 1940; Sanford 1961; Brogly 1984; Brogly and Martini 1990). Grey-green colouration occurs as scattered blebs, along joints and

| | **Northern Ontario** | | **Southern Ontario** | |
|---|---|---|---|---|
| **Series** | **Hudson Bay and Moose River basins** | **Lake Timiskaming outlier** | **South-Central and Southwestern Ontario** | **Eastern Ontario** |
| Upper Ordovician | Red Head Rapids Fm. | ? | Queenston Fm. | Queenston Fm. |
| | Churchill River Gp. | Dawson Point Fm. (Liskeard Gp.) | Georgian Bay Fm. | Carlsbad Fm. |
| | | | Blue Mountain Fm. | Billings Fm. |

**Figure 20.15.** Stratigraphic units of Sequence 3 (? indicates uncertain age).

**Table 20.3.** Late Ordovician (Maysvillian to Gamachian) units in Ontario—Depositional Sequence 3 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Southern Ontario** | | | | | | |
| Queenston Formation | Southwestern, south-central and eastern Ontario | Richmondian | Part of a large deltaic complex encompassing (in Ontario) nonmarine (?), marginal marine (coastal flood plain: intertidal to supratidal environs) and shallow marine environments | 45 to 335 m | Maroon, slightly calcareous to noncalcareous, sparsely fossiliferous and locally gypsiferous shale; interbeds of grey-green shale, bioclastic limestone and calcareous siltstone; grey-green colouration along some joints and bedding planes and in uppermost beds; wave and current ripples, desiccation cracks and horizontal laminae in siltstone and limestone interbeds.<br>Upper contact with Whirlpool or Manitoulin Fms. in southwestern and south-central Ontario is sharp and unconformable.<br>Lower contact is gradational with Georgian Bay or Carlsbad Fms. and marked by the base of the lowest significant bed of red shale. | Caley 1940;<br>Sanford 1961;<br>Brogly 1984;<br>Brogly and Martini 1990;<br>Liberty and Bolton 1971;<br>Williams 1991 |
| Georgian Bay Formation | Southwestern and south-central Ontario | Maysvillian to Richmondian | Represents a shallowing-upward storm-dominated shelf succession | 125 to 200 m (100 m avg.) | Blue-grey shale with minor siltstone and limestone interbeds; pale grey to cream, fossiliferous limestone and dolostone interbeds become more predominant upward and northward.<br>Graded beds; gutter casts; scour-and-fill structures; current- and wave-formed ripple marks; abundant body and trace fossils.<br>Lower contact with Blue Mountain Fm. is gradational and defined at last appearance of calcareous and silty hardbands >5 cm thick. | Liberty 1964;<br>Liberty and Bolton 1971;<br>Telford et al. 1981;<br>Horst 1978;<br>Byerley and Coniglio 1989;<br>Kerr and Eyles 1991 |
| Blue Mountain Formation | Southwestern and south-central Ontario | Maysvillian | Deposition below storm wave base | up to 60 m | Blue-grey, poorly fossiliferous, noncalcareous shale; fissile. In Toronto area, unit is predominantly brown to brown-grey shale (locally referred to as Rouge River Mb.).<br>Lower contact with Collingwood Mb. of Lindsay Fm. (Sequence 2) is sharp. Where Collingwood Mb. is absent, a phosphate lag marks the disconformable contact with the lower member of the Lindsay Fm. | Russell and Telford 1983;<br>Liberty 1969;<br>Williams 1991 |
| Carlsbad Formation | Eastern Ontario | Maysvillian to Richmondian | Storm-influenced shelf environment | up to 186 m | Interbedded, dark grey, calcareous and noncalcareous shale, fossiliferous siltstone and medium to pale grey, bioclastic limestone.<br>Cross-bedding; flute casts and ripple marks; moderately fossiliferous.<br>Lower contact with Billings Fm. is gradational. | Wilson 1946;<br>Williams 1991 |

**Table 20.3.** Late Ordovician (Maysvillian to Gamachian) units in Ontario—Depositional Sequence 3 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Billings Formation | Eastern Ontario | Maysvillian | Deposition below storm wave base | up to 62 m | Dark brown to black, slightly calcareous to noncalcareous shale with thin interbeds of dark grey limestone; pyritized fossils; moderately fossiliferous.<br>Lower contact with Eastview Mb. of Lindsay Fm. (Sequence 2) is conformable and defined on basis of carbonate content of the shale and limestone bed thickness. | Williams 1991;<br>Wilson 1946 |
| **Northern Ontario** | | | | | | |
| Red Head Rapids Formation | Hudson Bay and Moose River basins | Richmondian to Gamachian | Restricted shallow marine, evaporitic environment | 32 to 98 m | Light grey to tan, micro- to fine-crystalline, thin-bedded, laminated limestone and dolostone; evaporites common (anhydrite and salt); feldspathic interbeds in lower part; sparsely fossiliferous.<br>Upper contact is unconformable with Severn River Fm. (Sequence 4).<br>Lower contact is conformable with Churchill River Gp. and unconformable with Precambrian rocks. | Norris 1986;<br>Sanford 1987 |
| Churchill River Group | Hudson Bay and Moose River basins | Maysvillian to Richmondian | Open marine platform | 90 to 114 m | Light yellow to grey, microcrystalline to bioclastic, cherty limestone and dolostone; bituminous and dolomitic limestone locally present; fossils common to abundant.<br>Lower contact disconformable(?) with Bad Cache Rapids Gp. and Boas River Fm. | Sanford et al. 1968;<br>Sanford 1987;<br>Cumming 1975;<br>Norris 1986 |
| Dawson Point Formation (Liskeard Group) | Lake Timiskaming outlier | Maysvillian to Richmondian | Depositional environment similar to Queenston, Georgian Bay and Blue Mountain Fms. | 30 m | Upper part: thin, red shale unit.<br>Middle part: slightly calcareous shale with calcarenite and silty hardbands.<br>Lower part: blue- to olive-grey, noncalcareous shale.<br>Lower contact with Farr Fm. (Sequence 2) is disconformable. | Russell 1984 |

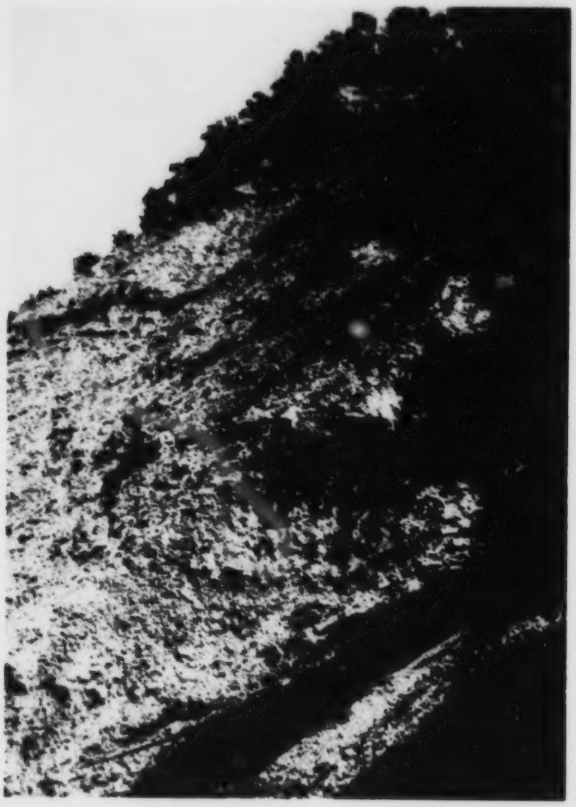

**Figure 20.16.** Blue-grey shales and interbedded limestones of the Georgian Bay Formation exposed at the Claybanks, on the shore of Georgian Bay, northwest of Meaford. At the top of this section, red shales of the lower part of the Queenston Formation are seen overlying the limestone beds at the top of the Georgian Bay Formation. (Photo: DA88–4–20)

**Figure 20.17.** Interbedded red shales and grey siltstones of the Queenston Formation, Bronte Creek, Oakville (photograph *courtesy of* G. McFall).

bedding planes, and within a thin interval beneath the sharp disconformable contact with the overlying Whirlpool or Manitoulin formations (*see* Sequence 4; Figure 20.18). The contact with the underlying Georgian Bay Formation is gradational, being placed at the base of the lowest significant red shale bed (*see* Figure 20.16).

On the Bruce Peninsula, the Queenston Formation contains grey limestone beds bounded by green shale (Armstrong 1989). These beds typically contain abundant bryozoans and brachiopods, and were identified as biostromes by Liberty and Bolton (1971). A stratigraphic hole drilled by the OGS near Wiarton (OGS–82–4, Figure 20.5; Johnson et al. 1985) intersected a 10 m section of this rock type approximately 10 m above the base of the Queenston Formation. This material appears very similar to the uppermost limestones and shales of the underlying Georgian Bay Formation, and may represent the initial lateral transition of the Queenston Formation into the upper part (Kagawong Member) of the Georgian Bay Formation of Manitoulin Island.

In the Hamilton area, the minor limestone and siltstone interbeds in the Queenston Formation contain ooids, brachiopods, ostracods, bryozoans, algal grains and vertical burrows (Brogly 1984). Sedimentary structures found in this unit include wave and current ripples, desiccation cracks and horizontal laminations (Brogly 1984).

On a regional scale, strata of the Queenston Formation form part of a large deltaic complex (encompassing fluvial, deltaic and shallow marine depositional environments), with sediments derived from the erosion of the Taconic Highlands. For the Queenston Formation strata in Ontario, a deltaic or terrestrial depositional environment was suggested by Grabau (1913) and Caley (1940) and an estuarine environment proposed by Foerste (1916). Recently, Brogly (1984) and Brogly and Martini (1990) interpreted a coastal flood-plain environment with supratidal and intertidal mud flat areas for exposures in the Hamilton area. Liberty and Bolton (1971) have interpreted a shallow marine environment for exposures on the Bruce Peninsula.

## EASTERN ONTARIO

In eastern Ontario, the Billings Formation (*see* Figures 20.8 and 20.15; Chart D, Map 2582, map case), consisting of shale (not organic rich) with thin interbeds of dark grey limestone, is considered to be approximately equivalent to the Blue Mountain Formation (Williams 1991). Lithologically, the Billings Formation seems most closely related to the Rouge River Member (of the Blue Mountain) which is best developed in the Toronto–Pickering area. The Billings Formation conformably overlies the Eastview Member of the Lindsay Formation and is distinguished from the Eastview Member by its noncalcareous nature (Williams 1991). This unit has a modest fauna of cephalopods, trilobites, brachiopods and graptolites, which are commonly pyritized (Williams 1991). Both the Blue Mountain and Billings formations are interpreted to have been deposited on a deep shelf environment below storm wave base (Williams 1991); clastic sediments were supplied by erosion of the Taconic Highlands in the southeast (Sanford, in press).

The Carlsbad Formation gradationally overlies the Billings Formation and is approximately correlative with the Georgian Bay Formation of south-central Ontario (*see*

**Figure 20.18.** Silurian dolostones of the Manitoulin Formation (Sequence 4) overlying Ordovician shales of the Queenston Formation (Sequence 3) at Indian Falls, north of Owen Sound. The uppermost 3 m of the typically red-coloured Queenston shales are green (middle of the section shown in photograph). (Photo: DA89–27–23)

Figure 20.15; Chart D, Map 2582, map case; Wilson 1946; Williams 1991). The similarities between the Carlsbad and Georgian Bay formations (rock type, thickness and fossil assemblage) suggests that the Frontenac Arch did not have a significant impact on sedimentation during the Late Ordovician.

Red to greenish-grey shales, siltstones and minor limestones of the Queenston Formation are found in the block faulted terrain of eastern Ontario, where it is the youngest preserved Paleozoic unit (Williams 1991; *see* Figure 4, Map 2582, map case, for distribution). The Queenston Formation conformably overlies the Carlsbad Formation in eastern Ontario and is lithologically similar to equivalent strata in south-central Ontario.

### NORTHERN ONTARIO

In the Lake Timiskaming outlier of northern Ontario (*see* Figure 20.1; Map 2544, map case), the unexposed Dawson Point Formation, consisting of shales and minor limestones (*see* Figure 20.15; Chart D, Map 2582, map case; Russell 1984) is correlated with the Blue Mountain, Georgian Bay and Queenston succession of southern Ontario on the basis of lithologic similarity (Russell 1984) and a graptolite assemblage (Barnes et al. 1981). This correlation suggests that the Upper Ordovician units now restricted to southern Ontario (Map 2544, map case) formerly covered much of central Ontario (Figure 4, Map 2582, map case).

In the Hudson Platform, Sequence 3 deposition is represented by the Churchill River Group and the Red Head Rapids Formation (*see* Figure 20.15; Chart D, Map 2582; Maps 2541 to 2543, map case). The Churchill River Group consists of open marine, platform carbonates, ranging from bioclastic to microcrystalline limestones and dolostones (Sanford et al. 1968; Cumming 1975; Sanford 1987). Bituminous crinoidal and dolomitic limestones are also present. The group consists of the lower Caution Creek Formation and the upper Chasm Creek Formation, which are, themselves, subdivided into members (not shown *in* Figure 20.15; Sanford et al. 1968; Cumming 1975).

The overlying Red Head Rapids Formation consists of limestones and dolostones (Cumming 1975; Norris 1986; Sanford 1987). In the Hudson Bay Basin, up to 20 m of halite and anhydrite have been reported in this unit. In the Moose River Basin, the formation is distinctly more clastic, although thin intervals of anhydrite and gypsum have also been reported (Norris 1986; Sanford 1987). The character and full extent of this evaporitic deposition is poorly known.

## Economic Geology

The fine-grained clastic rocks which characterize Sequence 3 in southern Ontario represent a significant resource for the clay industry. In the late 1980s, the demand for clay bricks in Ontario outstripped the domestic supply, particularly in the Toronto region, with the result that up to 100 million bricks were imported annually from the United States (Guillet and Joyce 1987). The Queenston Formation is the major source of raw material for the production of bricks in Ontario, owing to its consistency, chemistry and colour (Guillet 1977; Kwong et al. 1985). Unfortunately, areas of shallow overburden desirable for quarrying are located just to the east of the Niagara Escarpment, where conservation interests and urbanization pressures have constrained access to this resource.

The Georgian Bay Formation was a principal source of raw material for the manufacture of bricks, with 4 quarries active near Toronto in 1961 (Guillet and Joyce 1987). At the Don Valley Brickyard quarry, which is no longer in operation, a thickness of as much as 22.5 m of this formation had once been exposed (Guillet and Joyce 1987). The Georgian Bay Formation continues to have good potential in the Toronto area and elsewhere as a supply of raw materials for brick manufacture (Guillet 1967; Kwong et al. 1985; Guillet and Joyce 1987). The unit subcrops over a broad area, providing more extractive potential in less environmentally sensitive and less urbanized areas. However, these shales require additional processing to remove or selectively grind the silty hardbands.

The Blue Mountain Formation was locally a source of raw material for pottery (Guillet and Joyce 1987) and although not currently utilized for brick or other heavy clay products, it does represent a good potential resource (Guillet 1977). Thick Quaternary cover is the main limiting factor on the exploitation of the Blue Mountain shales.

# DEPOSITIONAL SEQUENCE 4: EARLY SILURIAN (EARLY TO MIDDLE LLANDOVERY)

## Overview

Depositional Sequence 4 is characterized by orogen-derived, fluvial to shallow marine clastic sediments, shed from the Taconic Highlands, which grade vertically and laterally into marine shales and carbonates. The lower contact of this sequence is an unconformity representing the

Late Ordovician regression and marking the Ordovician–Silurian system boundary. The upper contact of the sequence, in southern Ontario, is a regional angular unconformity beneath the basal units of Sequence 5, formed by uplift of the Algonquin Arch (*see* Figure 20.2) in the middle Llandovery. The Hudson Platform appears to have remained emergent throughout most of Sequence 4 time, with laterally extensive carbonate sediments and minor craton-derived clastic sediments deposited near the close of this sequence.

Subsequent to the Late Ordovician regression, which marked the close of Sequence 3, further Taconic tectonism caused transgression of the early Llandovery inland sea over a wide area of the southern Canadian Shield, extending northward to at least the Lake Timiskaming outlier area (Figure 5, Map 2582, map case; Bolton and Copeland 1972; Sanford, in press). The initial transgressive sediments of Sequence 4, consisting of orogen-derived fluvial to marginal marine clastic sediments, pass vertically into deltaic and shallow marine clastic sediments and minor carbonate sediments and laterally (northwestward) into carbonates and marine shales of the Michigan Basin succession (Chart D, Map 2582, map case; Fisher 1954; Amsden 1955; Bolton 1957; Winder and Sanford 1972; Sanford, in press). A lithologically similar and approximately equivalent succession of carbonate sediments and subordinate orogen-derived marine clastic sediments were also deposited by this transgression in the Lake Timiskaming outlier area (Chart D, Map 2582, map case; Poole et al. 1970; Russell 1984).

The absence of Sequence 4 sediments in eastern Ontario may be attributed to post–Sequence 5 subaerial erosion associated with epeirogenic uplift during the early to middle Ludlow Stage (Sanford et al. 1985). Alternatively, the absence of these strata may be attributed to late Paleozoic (*see* Sequence 9), Mesozoic or later erosion.

The Early Silurian marine transgression extended into northern Ontario, probably via the Williston Basin, but did not reach the Hudson Platform until the middle Llandovery, near the end of this depositional sequence (Chart D, Map 2582, map case; Sanford 1987). Sequence 4 carbonate and minor craton-derived clastic strata are widespread in the Hudson Bay and Moose River basins and probably extended farther south onto the Canadian Shield than they do at present (Figure 5, Map 2582, map case). Much of these deposits were removed, along with older Sequence 3 strata and younger Sequence 5 strata, as a result of uplift and erosion that probably began along the northern margins of the Fraserdale Arch in early Ludlow time (Sanford 1987).

Lithologic details of Sequence 4 units are shown in Table 20.4. Stratigraphic nomenclature is shown on Figure 20.19. The present distribution of these units is shown on Maps 2541 to 2544 (map case). The pre-erosion maximum extent is shown on Figure 5, Map 2582 (map case). Distribution of the sequence is shown on Maps 2575 to 2578 (map case). In the following section, details of the stratigraphy of Sequence 4 are presented, subdivided on a geographic basis.

# Geology

## SILURIAN TERMINOLOGY

The Silurian is subdivided into 2 time stratigraphic intervals, Early and Late, which correspond to the rock stratigraphic terms Lower and Upper Silurian, respectively. However, in the northeastern United States and adjacent areas in Canada, where the New York succession and its fauna have been well studied, a threefold rock stratigraphic subdivision of Lower, Middle and Upper Silurian has evolved (*see* Figure 20.3; Chart D, Map 2582, map case; Berry and Boucot 1970). These terms, approximately correlative with the North American series terms Alexandrian, Niagaran and Cayugan, are defined primarily on a lithostratigraphic basis, owing to the relatively unfossiliferous nature of the Lower and Upper Silurian strata. Therefore, Berry and Boucot (1970) recommended against extrapolation of these rock stratigraphic terms outside New York State and adjacent areas. As shown in Figure 20.3 and on Chart D (Map 2582, map case), sequences 4, 5 and 6 approximately correspond to the rock stratigraphic terms Lower, Middle and Upper Silurian, respectively.

## SOUTHERN ONTARIO

The Silurian strata of southern Ontario are perhaps the best understood rocks in the region. This is due, in part, to their excellent exposure along the Niagara Escarpment (*see* Figure 20.1) and to the considerable oil and gas resources within several Silurian units in the subsurface (Sanford 1969; Bailey and Cochrane 1986).

The Niagara Escarpment (*see* Figures 20.1 and 20.20) is a prominent erosional cuesta geographically separating lowlands, underlain by Ordovician strata, in the east, from uplands of Silurian and Devonian strata in the west (Hewitt 1971). The escarpment can be traced from near Rochester, New York, to Niagara Falls, northwestward across southern Ontario to Manitoulin Island and thence to northern Michigan. Thus, the escarpment exposes Lower and Middle Silurian strata of the Appalachian Basin, the Algonquin Arch area and the Michigan Basin (*see* Figure 20.2), and displays the lateral variations in rock type and facies among these various depositional settings.

The Silurian section of the Niagara Escarpment (both sequences 4 and 5) has been long studied, the first comprehensive account having been provided by Bolton (1957). A regional study of Silurian strata in outcrop and the subsurface of southwestern Ontario was made by Sanford (1969), who also identified lithofacies patterns in the Middle Silurian carbonate rocks (*see* Sequence 5). A more recent summary of the Niagara Escarpment succession in southern Ontario (Telford 1978) was based on mapping by Liberty, Feenstra and Telford (1976a, 1976b), Liberty, Bond and Telford (1976a, 1976b, 1976c), Bond et al. (1976), Bond and Telford (1976) and Telford (1976a, 1976b). New lithofacies maps of the Silurian on Manitoulin Island (Johnson and Telford 1985a–f), Cockburn Island (Wolf 1986) and the Bruce Peninsula (Armstrong and Dubord 1992; Armstrong, in prep. b, in prep. c) provide recent data on the northern part of the escarpment. The recent application of new depositional models to detailed outcrop and core studies has lead to proposed stratigraphic revisions for part of the Silurian succession in the Niagara Peninsula and western New York State (Duke 1987, 1991; Fawcett et al. 1988; Brett et al. 1990, 1991).

In southern Ontario, Sequence 4 strata have been assigned to the Cataract Group (*see* Figure 20.19). On the Niagara Peninsula, the term Medina Group has been applied by some workers to this succession. Some workers also include the overlying Thorold Formation (*see* Sequence 5; e.g., Kilgour 1972). There is some uncertainty associated with the assignment of some units to either the Cataract or the overlying Clinton Group (*see* Sequence 5) (cf., Bolton 1957; Liberty and Bolton 1971). This may be due to the lithostratigraphic application of biostratigraphically based terms and is further discussed in this section and in Sequence 5.

Forming the base of Sequence 4 in southern Ontario are the sheet-like sandstones of the Whirlpool Formation (Figure 20.21). This unit outcrops along the Niagara Escarpment, from the Niagara Gorge to near Collingwood, and extends westward in the subsurface to the Long Point area of Lake Erie (Sanford 1969; Middleton et al. 1987; Figure 5, Map 2582, map case). Previously considered a shallow marine or eolian deposit (e.g., Grabau 1913; Fisher 1954; Bolton 1957; Martini 1971), recent sedimentological studies have proposed that the Whirlpool Formation consists of a lower fluvial unit and an upper marine unit (Salas 1983; Middleton et al. 1983, 1987; Middleton and Rutka 1986; Rutka 1986; Rutka et al. 1991). Sedimentary structures (*see* Table 20.4), paleocurrent indicators and the absence of marine fauna or traces indicate a braided fluvial

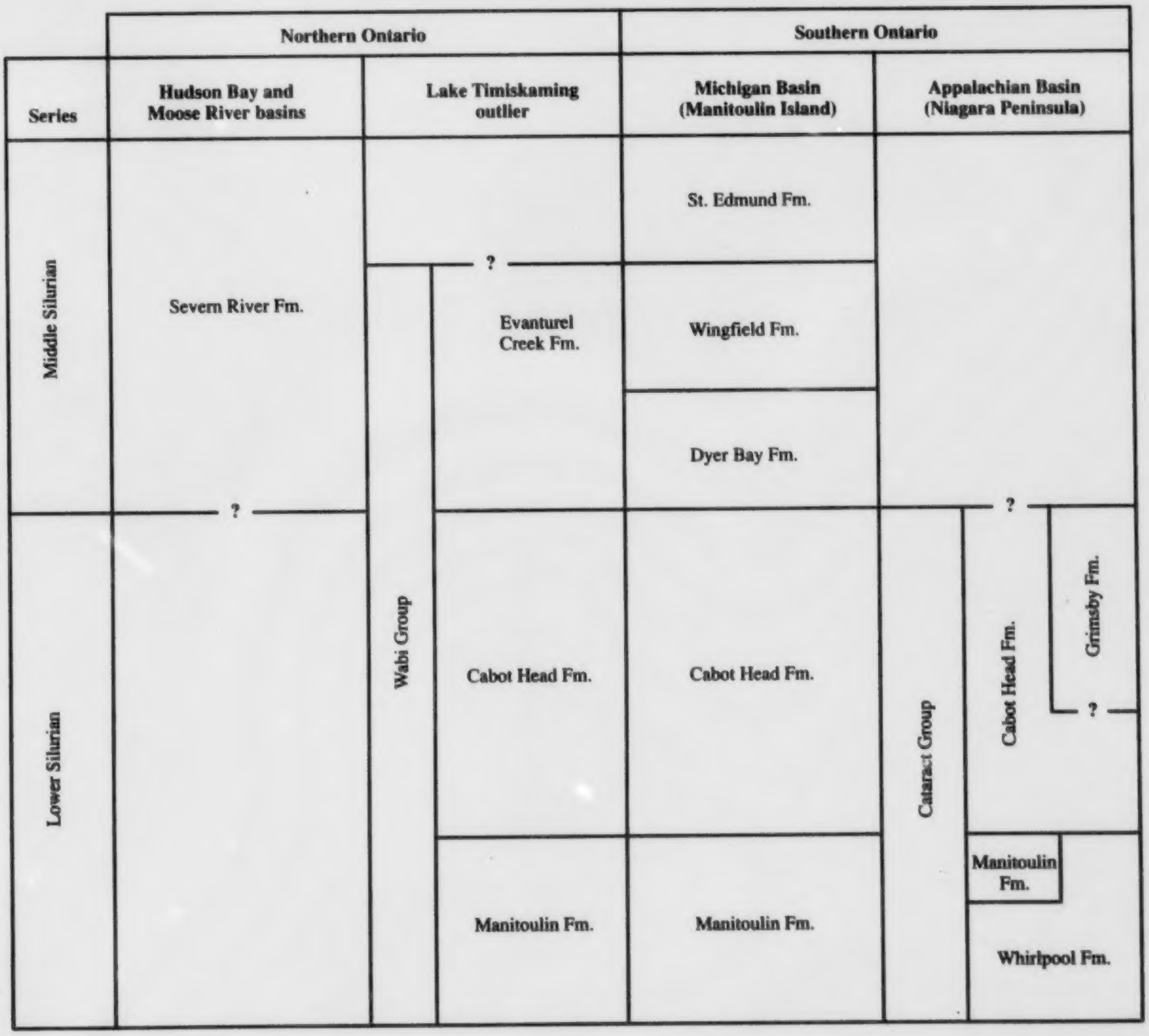
Northern Ontario
Southern Ontario
Series
Hudson Bay and Moose River basins
Lake Timiskaming outlier
Michigan Basin (Manitoulin Island)
Appalachian Basin (Niagara Peninsula)
Middle Silurian
Lower Silurian
Severn River Fm.
?
?
Wabi Group
Evanturel Creek Fm.
Cabot Head Fm.
Manitoulin Fm.
St. Edmund Fm.
Wingfield Fm.
Dyer Bay Fm.
Cabot Head Fm.
Manitoulin Fm.
?
Cataract Group
Cabot Head Fm.
Grimsby Fm.
?
Manitoulin Fm.
Whirlpool Fm.

**Figure 20.19.** Stratigraphic units of Sequence 4 (? indicates uncertain age).

Table 20.4. Early Silurian (early to middle Llandovery) units in Ontario—Depositional Sequence 4 (described according to geographic location and in order of youngest to oldest unit)

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Southern Ontario** | | | | | | |
| St. Edmund Formation | Central Ontario (North Channel Islands and Bruce Peninsula) | Llandovery | Intertidal to supratidal (facies 1) and deeper intertidal to subtidal (facies 2) environments | up to 25 m | Two facies: 1) occurs primarily in upper part and consists of cream-buff, microcrystalline (lithographic-textured), thin-bedded dolostones with desiccation cracks; and 2) buff to dark brown, fine- to medium-crystalline, thick-bedded dolostones; large stromatoporoids locally abundant.<br>Lower contact is sharp with the Wingfield Fm. | Johnson and Telford 1985c; Johnson and Telford 1981a, 1981b; Liberty and Bolton 1971; Armstrong 1989; Armstrong and Goodman 1990; Copper 1978 |
| Wingfield Formation | Central Ontario (North Channel Islands and Bruce Peninsula) | Llandovery | Shallow subtidal, restricted marine environment | 2 to 15 m (12 m avg.) | Interbedded olive-green, non- to slightly calcareous, sparsely fossiliferous shales and thin-bedded brown, argillaceous dolostones. Planar and ripple cross laminae (dolostones); mudcracks (shales).<br>Lower contact is sharp to gradational with Dyer Bay Fm. | Liberty and Bolton 1971; Bolton 1957; Johnson and Telford 1981a, 1981b; Armstrong 1989; Armstrong and Goodman 1990; Copper 1978 |
| Dyer Bay Formation | Central Ontario (North Channel Islands and Bruce Peninsula + subsurface of southwestern Ontario) | Llandovery | Shallow subtidal environment | up to 7.6 m (2 to 4 m avg.) | Grey to brown, thin- to medium-bedded, argillaceous and bioturbated dolostones; fossiliferous and locally (Manitoulin Is.) biostromal (coral and bryozoa) with shell beds (brachiopods); green-grey shaly partings; soft-sediment deformation, ripple marks; horizontal laminae; fossiliferous; bioturbated.<br>Lower contact: conformable, yet sharp and well defined with Cabot Head Fm. | Liberty and Bolton 1971; Johnson and Telford 1985f; Armstrong 1988; Armstrong and Goodman 1990; Copper 1978 |
| Grimsby Formation (Cataract Group) | Southwestern Ontario (Niagara Peninsula + subsurface) | Llandovery | Episodic progradation of tide-dominated (tidal channels), wave-influenced shoreline into a storm-dominated shallow marine basin | up to 15.8 m | Upper part: interbedded, maroon shales and lenticular to channelled, red and white to yellow, mottled sandstones.<br>Lower part: maroon shales with interbeds of fine-grained hummocky to planar cross-stratified sandstone.<br>Graded beds, wave ripples, hummocky, planar and trough cross-bedding; desiccation cracks; soft-sediment deformation; body and trace fossils.<br>Upper contact: sharp erosional and disconformable with Thorold Fm. (Sequence 5).<br>Lower contact: gradational with Cabot Head Fm. | Sanford 1969; Martini 1971; Fisher 1954; Bolton 1957; Brett and Calkin 1987; Duke and Brusse 1987; Friedman et al. 1982; Duke et al. 1991; Duke 1991 |
| Cabot Head Formation (Cataract Group) | Southwestern Ontario and northern Ontario (Lake Timiskaming outlier) | Llandovery | Overall shallowing upward from storm-influenced, offshore basinal to marginal marine environment | 10 to 39 m | Southern Ontario: maroon, to green and grey, noncalcareous shales with minor calcareous sandstone, dolostone and limestone interbeds; generally unfossiliferous, but with abundant bryozoa in some calcareous beds.<br>Lake Timiskaming outlier: dolomitic, red shale, brecciated and interbedded with green shale, green-grey sandstone and gypsum.<br>Lower contact: gradational with Manitoulin or Whirlpool Fms. | Bolton 1957; Russell 1984; Duke 1982; Duke and Brusse 1987; Sanford 1969 |

**Table 20.4.** Early Silurian (early to middle Llandovery) units in Ontario—Depositional Sequence 4 (described according to geographic location and in order of youngest to oldest unit)

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Manitoulin Formation (Cataract Group) | Southwestern Ontario and northern Ontario (Lake Timiskaming outlier) | Llandovery | Storm-dominated, shallow shelf; near-shore to offshore, patch, barrier reef and interreef environments, occasionally affected by storms (Manitoulin Island) | up to 25 m | Southwestern Ontario: grey, thin- to medium-bedded, moderately fossiliferous dolostone; fine- to medium-crystalline; weathers buff to blue-grey.<br>Manitoulin Island (eastern part) and northern Bruce Peninsula: arcuate and ovoid-shaped bioherms.<br>Lake Timiskaming outlier: brownish-buff, very fine- to fine-crystalline dolostone with shaly partings and bioclastic layers.<br>Moderately fossiliferous (some fossils silicified); trough cross-bedding; laminated and graded?; low angle cross-stratification; bioturbation.<br>Lower contact: conformable with Whirlpool Fm. (southwestern Ontario) and sharp and disconformable with Queenston (southern Ontario) and Dawson Point (Lake Timiskaming outlier) Fms. (both Sequence 3). | Russell 1984;<br>Copper 1978;<br>Copper and Fay 1989;<br>Johnson and Telford 1985a-f;<br>Bolton 1957, 1964;<br>Leggitt 1985;<br>Anastas and Coniglio 1991;<br>Grawberger 1978 |
| Whirlpool Formation (Cataract Group) | Southwestern Ontario (Niagara Peninsula to Collingwood + subsurface) | Llandovery | Shallow marine (upper unit); braided fluvial (lower unit) | up to 9 m | White to light grey, maroon, fine-grained, orthoquartzitic sandstone.<br>Upper unit: symmetrical ripple marks, marine body and trace fossils; hummocky cross-stratification.<br>Lower unit: unfossiliferous, planar and trough cross-bedding, current ripple cross-laminations, channels, plane beds, soft-sediment deformation and shale intraclasts.<br>Lower contact: sharp and unconformable with the Queenston Fm. (Sequence 3). | Fisher 1954;<br>Bolton 1957;<br>Martini 1971;<br>Middleton et al. 1983, 1987; Rutka 1986;<br>Salas 1983 |
| **Northern Ontario** | | | | | | |
| Severn River Formation | Hudson Bay and Moose River basins | Llandovery | Undetermined | 45 to 248 m | Brown to tan, fine-crystalline limestone and dolostone with local bioclastic limestone and intraformational conglomerate beds. Sparsely fossiliferous (brachiopods); bioturbated.<br>Basal beds are sandy and conglomeratic where the Severn River Fm. directly overlies Precambrian rocks.<br>Lower contact with underlying Red Head Rapids Fm. (Sequence 3) is unconformable. | Sanford et al. 1968;<br>Norris 1986 |
| Evanturel Creek Formation (Wabi Group) | Lake Timiskaming outlier | Llandovery | Undetermined | up to 15 m | Green and buff, fine- to medium-crystalline dolostone and green mudstones and shales with minor interbeds of intraformational, shale-clast conglomerate. Mudcracks and ripple marks.<br>Upper contact with Earlton Fm. (Sequence 5) is disconformable; lower contact conformable with Cobalt Head Fm. | Russell 1984 |
| Cabot Head Formation (Cataract Group) | Lake Timiskaming outlier | | | | *see* description under Southern Ontario | |
| Manitoulin Formation (Cataract Group) | Lake Timiskaming outlier | | | | *see* description under Southern Ontario | |

**Figure 20.20.** View, to the northeast, of the Niagara Escarpment on the eastern side of the Bruce Peninsula. The white cliffs in the foreground, consisting of Silurian dolostones of the Amabel Formation (Sequence 5), overlook Cape Croker in the distance, which is underlain by Ordovician strata (Sequence 3). (Photo: DA88–11–30)

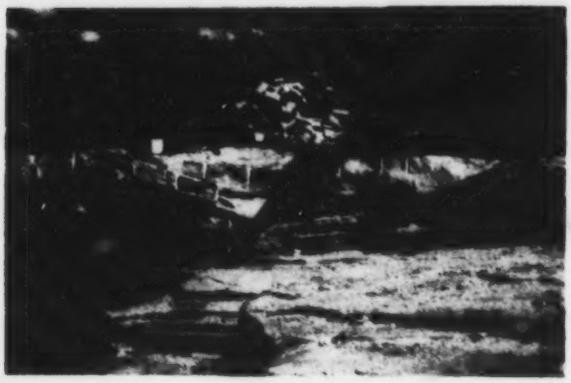

**Figure 20.21.** Lower part of the Whirlpool Formation in the Rice and McHarg Quarry at Limehouse, west of Georgetown. Photograph shows part of a 6 m wide channel-like scour (infilled with trough cross-bedding) cutting into horizontally laminated sandstone (scale bar in centre is divided into 10 cm intervals; photograph *courtesy of* M. Rutka).

depositional environment for the lower unit. The upper unit is clearly shallow marine in origin, consisting mainly of interbedded symmetrically rippled sandstone and shale, with many beds containing marine trace and body fossils (Rutka 1986; Middleton et al. 1987).

The lower contact of the Whirlpool Formation with the underlying Ordovician Queenston Formation shales is a knife sharp unconformity, probably representing the entire Gamachian stage (*see* Sequence 3; Barnes et al. 1981). The upper contact is gradational with either shales of the Cabot Head Formation (east of Stoney Creek) or dolostones of the Manitoulin Formation (west of Stoney Creek) (Chart D, Map 2582, map case).

Dolostones of the Manitoulin Formation gradationally overlie the sandstones of the Whirlpool Formation north and west of Stoney Creek (Niagara Peninsula). The Whirlpool Formation also appears to give way laterally to the Manitoulin Formation. These 2 units may be, in part, time equivalent (Chart D, Map 2582, map case). Beyond the pinch-out of the Whirlpool Formation (i.e., northwest of Collingwood), the Manitoulin Formation sharply and disconformably overlies the Upper Ordovician Queenston Formation (of Sequence 3). In addition to outcropping along the Niagara Escarpment, the Manitoulin Formation occurs extensively in the subsurface of southwestern Ontario and Michigan (Sanford 1969) and is clearly identifiable in the Lake Timiskaming outlier (Russell 1984), suggesting a relatively widespread, pre-erosion distribution (Figure 5, Map 2582, map case).

Generally, the Manitoulin Formation consists of moderately fossiliferous dolostone (Bolton 1957, 1964; Liberty and Bolton 1971). Some fossils are silicified and well preserved, however, dolomitization has obscured most primary structures and locally only fossil "ghosts" remain. On weathered surfaces, depositional features such as cross-bedding and burrows can be discerned (Armstrong 1988, 1989).

On the eastern end of Manitoulin Island, bioherms composed of corals, stromatoporoids, bryozoans and crinozoans occur primarily in the upper part of the Manitoulin Formation (Copper 1978; Copper and Fay 1989). These are among the earliest and largest Silurian bioherms in North America. The largest of these, located near Manitowaning, is an arcuate "barrier" reef, up to 4 km long by 200 m wide, which is founded on the disconformable basal contact of the formation (Grawbarger 1978; Johnson and Telford 1985f; Copper and Fay 1989). At Cabot Head, on the northeastern end of the Bruce Peninsula, small bryozoan-dominated bioherms occur near the base of the Manitoulin Formation (Anastas and Coniglio 1991).

The fauna and sedimentary structures (*see* Table 20.4) of the Manitoulin Formation are consistent with a shallow, storm-influenced shelf setting (Leggitt 1985; Anastas and Coniglio 1991). Biohermal development in the Manitoulin Island area indicates a shallower environment there (Copper and Fay 1989).

The Manitoulin Formation, or where it is absent, the Whirlpool Formation, is overlain by shales of the Cabot Head Formation. These shales occur throughout southwestern Ontario (Sanford 1969) and have been identified as far north as the Lake Timiskaming outlier (Figure 5, Map 2582, map case; Russell 1984). The Cabot Head Formation in the eastern part of the Niagara Peninsula (Power Glen Formation of Bolton 1957), is stratigraphically equivalent to the lower part of the Cabot Head Formation to the northwest (Chart D, Map 2582, map case; *see* Figure 20.19). The upper part of the Cabot Head Formation to the northwest is laterally equivalent to the Grimsby Formation of the Niagara Peninsula (Bolton 1957).

The Cabot Head Formation shales contain subordinate calcareous sandstone, limestone and dolostone interbeds (Sanford 1969). The unit is generally unfossiliferous, although abundant bryozoans occur in some calcareous beds.

The lower contact of the Cabot Head Formation with the Manitoulin Formation is everywhere gradational. On the

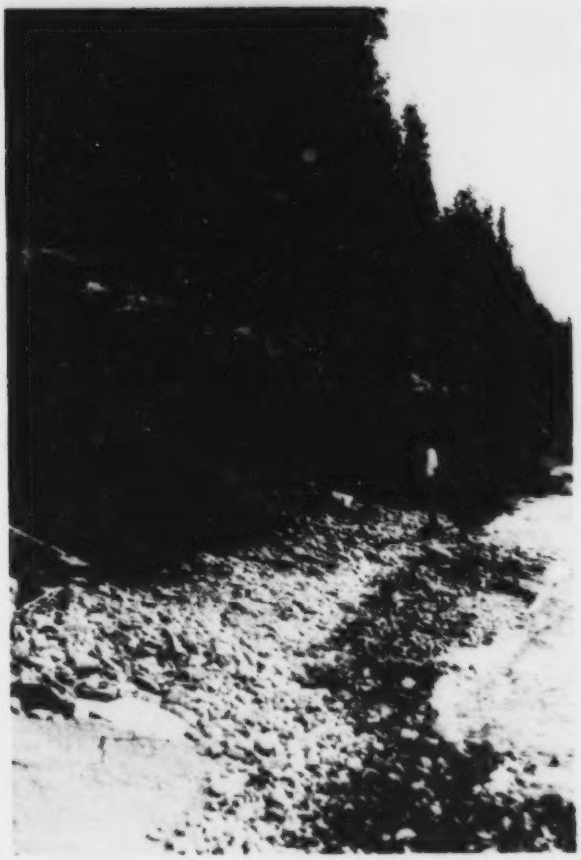

**Figure 20.22.** Thin-bedded dolostones of the Dyer Bay Formation overlying the red (blue-green near the contact) shales of the Cabot Head Formation, near Cape Chin on the Bruce Peninsula. (Photo: DA87–8–6)

Niagara Peninsula, its upper contact with the red Grimsby Formation is more gradational than the lower contact and is generally drawn at the colour change from grey to red (Bolton 1957). This colour change criterion has been recently questioned by Duke (1987) and Duke et al. (in press), leading some workers (e.g., Brett et al. 1991) to move the base of the Grimsby Formation down into the Cabot Head Formation. To the north and west, the Cabot Head is disconformably overlain by carbonate rocks of Sequence 5 (Reynales and Fossil Hill formations). On Manitoulin Island and the Bruce Peninsula, its upper contact with the Dyer Bay Formation appears conformable, yet sharp and well defined (Figure 20.22).

Sand to shale ratios indicate that the bulk of the clastic material of the Cabot Head formation is derived from the Taconic Highlands to the southeast (Sanford 1969). However, towards its northern limit, a locally significant craton-derived component is likely. In the Niagara Peninsula area, the Cabot Head Formation was deposited in a shallowing upward setting, ranging from storm-influenced (Duke 1982) offshore to marginal marine towards its top (Brett et al. 1990). In the Manitoulin area, a nearshore to marginal marine depositional setting is reported (Copper 1978). In subsurface of southwestern Ontario, a more basinal offshore setting is indicated for the Cabot Head Formation by its grey colouration (Sanford 1969).

The maroon sandstones and shales of the Grimsby Formation conformably overlie the Cabot Head Formation (Power Glen Formation of Bolton 1957) on the Niagara Peninsula and grade laterally into the upper part of the Cabot Head Formation to the northwest (Chart D, Map 2582, map case; Sanford 1969). The Grimsby Formation thins northwestward and has not been identified north of Hamilton (Fisher 1954; Bolton 1957; Martini 1971). This formation can be grossly subdivided into a lower, shale-dominated unit with sandstone interbeds and an upper unit consisting of sandstone with intraformational conglomerate and sandy shale interbeds (Bolton 1957; Kilgour and Liberty 1981; Brett and Calkin 1987; Duke and Brusse 1987).

Hematite-coated sand grains and other sedimentological features in the Grimsby Formation suggest a variety of shallow water to marginal marine environments in a prograding deltaic setting (Martini 1971; Friedman et al. 1982). The coarsening upward nature of the Grimsby Formation is consistent with recent interpretations (Duke et al., in press), which suggest that the Grimsby, as well as the Cabot Head and possibly the Thorold formations, represent part of a shallow marine, storm-dominated, prograding shoreline succession.

The upper contact of the Grimsby Formation with the Thorold Formation (*see* Sequence 5) is usually drawn at the abrupt colour change from the red of the Grimsby Formation to the green of the Thorold Formation (Bolton 1957; Kilgour 1972). However, recent studies indicate that colour variation here may be diagenetic and perhaps not adequate for defining formational boundaries (Duke 1987; Fawcett et al. 1988). Locally, however, this contact also exhibits reworked Grimsby sediments in the lowermost Thorold Formation (Fisher 1954; Bolton 1957; Brett and Calkin 1987).

Overlying the Cabot Head Formation shales on the Bruce Peninsula and on the islands of the North Channel of Lake Huron are a succession of carbonates and shales termed (in ascending order) the Dyer Bay, Wingfield and St. Edmund formations (*see* Figure 20.19; Chart D, Map 2582, map case). All 3 units are extensively exposed on Manitoulin Island (Johnson and Telford 1985a–f). On the Bruce Peninsula, these units are progressively truncated southward by a regional angular unconformity at the base of the overlying Fossil Hill Formation of Sequence 5 (Chart D, Map 2582, map case; Armstrong and Goodman 1990). The angular nature of this unconformity indicates that cratonic uplift centred on the Algonquin Arch (*see* Figure 20.2) area peneplaned these Sequence 4 sediments prior to Sequence 5 deposition. The angular nature of the corresponding regional unconformity, which should be present on the south side of the Algonquin Arch at this stratigraphic position (base of Thorold and Reynales formations of Sequence 5), is not obvious due to the lack of significant marker beds in the Cabot Head Formation. The southern extent of the Dyer Bay, Wingfield and St. Edmunds formations prior to this erosion is unknown, although they may have originally been deposited over the Algonquin Arch and, therefore, be laterally equivalent to the upper Cabot Head Formation (and Grimsby?) strata of the Niagara Peninsula (*see* Figure 20.19; Chart D, Map 2582, map case; Sanford 1969; Poole et al. 1970, Chart II).

Biostratigraphic evidence (Barnes et al. 1978; Copper 1978; Brett et al. 1990) suggests that the Dyer Bay to St. Edmund succession may be equivalent to the Neahga and Reynales formations of New York State and the Niagara Peninsula, and thus should be included in the overlying Sequence 5. The units are here included with Sequence 4 because of the well-developed regional unconformity at the top of this succession on the Bruce Peninsula (Armstrong 1989; Armstrong and Goodman 1990) and the apparent lack of unconformity between the Dyer Bay and Cabot Head formations.

The argillaceous dolostones of the Dyer Bay Formation (*see* Table 20.4) are generally fossiliferous and, locally, biostromal. On Manitoulin Island, the unit contains rare coral and bryozoan biostromes and shell beds rich in the pentamerid brachiopod *Virgiana* (Copper 1978). The Dyer Bay Formation is commonly bioturbated and displays a variety of sedimentary structures including ripple marks and horizontal lamination (Armstrong 1988), which suggest deposition in a shallow subtidal environment (Copper 1978). The unit is best exposed at the Ice Lake roadcut on Manitoulin Island (Telford et al. 1981; Johnson and Telford 1985c) and at its type locality on the Bruce Peninsula (Liberty and Bolton 1971). In its outcrop belt, the Dyer Bay Formation is continuous southward to Owen Sound (Bolton 1957; Armstrong 1989), commonly forming a subordinate escarpment beneath the main Niagara Escarpment. In the subsurface, this unit has been reported below Essex and Kent counties of southwestern Ontario (Sanford 1969).

The Wingfield Formation consists of interbedded shales and thin-bedded dolostones. The dolostones locally exhibit planar laminations and the shales exhibit mudcracks (*see* Table 20.4). The contact of this unit with the underlying Dyer Bay Formation is sharp (Bolton 1957; Liberty and Bolton 1971) to gradational (e.g., Johnson and Telford 1981a, 1981b). The Wingfield Formation is only present in outcrop as far south as Colpoy's Bay on the Bruce Peninsula (Armstrong 1989). The depositional environment of the Wingfield Formation appears to have been shallower and more restricted than that of the Dyer Bay Formation (Copper 1978).

The overlying dolostones of the St. Edmund Formation consist of 2 lithofacies: an upper microcrystalline lithofacies (Mindemoya facies, cf., Mindemoya Formation of Liberty 1968) and a lower coarser crystalline lithofacies. The microcrystalline strata display desiccation cracks, suggesting deposition in an intertidal to supratidal environment (Copper 1978). The coarser lithofacies, containing the so-called "oncolite" beds which consist of large bulbous stromatoporoids (Johnson and Telford 1985c), was probably deposited in a deeper intertidal to subtidal environment.

The St. Edmund Formation thins southward, from a maximum of about 25 m on Manitoulin Island (Johnson and Telford 1981a, 1981b) to about 3 m on the northern Bruce Peninsula (Liberty and Bolton 1971) and is absent farther south at Lion's Head. Subsurface records on Manitoulin Island indicate that the contact with the underlying Wingfield Formation is sharp (Johnson and Telford 1981a, 1981b).

### NORTHERN ONTARIO

In the Lake Timiskaming outlier (*see* Figure 20.1; Map 2542, map case), Sequence 4 is represented by the Wabi Group (Chart D, Map 2582, map case; *see* Figure 20.19), which consists of (in ascending order) the Manitoulin, Cabot Head and Evanturel Creek formations (Russell 1984). Only the Evanturel Creek is exposed in the outlier; the other 2 units being encountered in OGS drill hole LT–1 (Russell 1984). In the outlier, both the Manitoulin and Cabot Head formations are very similar in thickness and lithologic character to these same units on Manitoulin Island and in southern Ontario (*see* Table 20.4).

The Evanturel Creek Formation consists predominantly of dolostones and shales with minor intraformational, shale-clast conglomerates (Russell 1984). This unit is lithologically similar to the Wingfield and Dyer Bay strata on Manitoulin Island and the Bruce Peninsula, and is likely approximately equivalent to these units (Chart D, Map 2582, map case).

On the Hudson Platform, Silurian deposition commenced in middle Llandovery time, near the close of this depositional sequence (Chart D, Map 2582, map case), with the deposition of carbonate sediments and minor craton-derived clastic sediments of the Severn River Formation. This unit was deposited in both the Moose River and Hudson Bay basins, which suggests that the Cape Henrietta Maria Arch was not a significant barrier during this time (Figure 5, Map 2582, map case). The Severn River Formation consists of fine-crystalline limestone and dolostone with local bioclastic and intraformational conglomerate beds (Sanford et al. 1968). Thin basal clastic beds are developed where the formation directly overlies Precambrian terrain. The presence of the pentamerid brachiopod *Virgiana* within the Severn River Formation suggests a biostratigraphic correlation with the Dyer Bay Formation of southern Ontario and the Wabi Group of the Lake Timiskaming outlier (Sanford et al. 1968).

## Economic Geology

The economic resources of Sequence 4 strata include building stone from the Whirlpool and Grimsby Formation sandstones, aggregates from the Manitoulin and St. Edmund Formation dolostones and natural gas from the "Clinton Sands".

The Whirlpool Formation was once a major source of building stone for the Toronto area (Hewitt 1964). Production from this unit is now limited to a couple of small quarries in the Orangeville area of the Niagara Escarpment. Of the various building stone colours present in the

Whirlpool Formation, the most prized has been the maroon "Credit Valley Stone" (Parks 1912), used for the Provincial Parliament buildings in Toronto. At present, the small DeForest Brothers Quarry, at the Forks of the Credit, is the only producer of this maroon Whirlpool Formation stone in Ontario (Ministry of Northern Development and Mines 1990a). The grey variety of this formation is still produced by several other small quarries (Ministry of Northern Development and Mines 1990b).

The Manitoulin Formation has good potential for aggregate resource development and is actively quarried in St. Vincent Township near Collingwood (Ontario Geological Survey 1992), and in Sarawak Township near Owen Sound (Armstrong 1989). The thin- to medium-bedded strata of this unit are quarried for landscaping and building stone at the Hercules Ornamental stone quarry near Foxey on Manitoulin Island (Johnson and Telford 1985b; Derry, Michener, Booth and Wahl and OGS 1989b). The St. Edmund Formation also has aggregate potential and has been quarried on Manitoulin Island (Johnson and Telford 1985c; Derry, Michener, Booth and Wahl and OGS 1989b). Several other units in the sequence (i.e., Dyer Bay, Evanturel Creek and Severn River formations) may also have limited aggregate potential.

The hydrocarbon resources of Sequence 4 strata are a significant component to Ontario's total production. Hydrocarbons are found in the Whirlpool Formation and, more significantly, in the sandstone beds of the upper Grimsby Formation. From an exploration and production viewpoint, these units and the overlying Thorold Formation (*see* Sequence 5) are collectively referred to as the "Clinton Sands" (Bailey and Cochrane 1986). The trapping mechanism for hydrocarbons within these rocks involves the pinch-out of higher porosity-permeability zones within sandstone to dolostone intervals.

In 1989, no oil was produced in Ontario from "Clinton–Cataract" strata. Total gas production from these units amounted to 274 179 thousand $m^3$, or 56% of the total Ontario production, with the majority coming from offshore in Lake Erie (Carter 1990). Cumulative oil production from these strata through to the end of 1989 amounted to just 7000 $m^3$, or 0.07% of the total Ontario production, whereas cumulative gas production was 10 497 000 thousand $m^3$, or 37% of total Ontario gas production (Carter 1990). Oil production is entirely from onshore wells, as oil exploration and extraction is not allowed offshore in Lake Erie.

## DEPOSITIONAL SEQUENCE 5: EARLY SILURIAN (MIDDLE LLANDOVERY) TO LATE SILURIAN (MIDDLE LUDLOW)

### Overview

In most of both northern and southern Ontario, Sequence 5 is characterized by extensive carbonate deposition. Mixed carbonate and orogen-derived clastic sedimentation was almost exclusively restricted to the Appalachian Basin during this sequence. Renewed cratonic uplift in the early to middle Ludlow(?), which caused local unconformity development and a significant change in sedimentation style and distribution, terminated Sequence 5.

The deposition of Sequence 5 sediments followed a depositional hiatus associated with a regional regression during the middle Llandovery. This regression affected all of southern Ontario, possibly as far north as the Lake Timiskaming outlier area (Russell 1984). In southern Ontario, uplift of the Algonquin Arch during this regression and erosion of Sequence 4 sediments resulted in the development of regional angular unconformities, with progressively younger units subcropping beneath the unconformity away from the arch (*see* Sequence 4).

Carbonates and orogen-derived marine clastic sediments of Sequence 5 were deposited in the Western St.Lawrence Platform (Cohee 1948; Colton 1970; Sanford 1969), blanketing southern Ontario and extending north to the Lake Timiskaming outlier area (Figure 6, Map 2582, map case; Russell 1984). They were likely also deposited in the Ottawa Embayment and Quebec Basin (both of the Central St. Lawrence Platform), but subsequently removed by erosion (*see* Sequence 9 for discussion of erosion; Sanford, in press).

In the southern part of the Appalachian Basin, sedimentary rocks of Sequence 5 are dominated by orogen-derived clastic rocks (Figure 6, Map 2582, map case; Cohee 1948; Beardsley and Cable 1983). Along the northern margin of the basin (including the Niagara Peninsula area), these clastic rocks intertongue with shelf carbonate rocks (Cohee 1948; Bolton 1957), which in turn intertongue, in southwestern Ontario (Bolton 1957; Sanford 1969; Winder and Sanford 1972) and adjacent Michigan (Cohee 1948), with complex reefal carbonate rocks of the Michigan Basin (Figure 6, Map 2582, map case). Thick barrier reef and carbonate bank deposits (Guelph and Amabel formations) formed around the Michigan Basin and, in response to accelerated basin subsidence, patch and pinnacle reefs developed along the inner (i.e., basinward) margins of the barrier complex in Ontario and Michigan (Figure 20.23; Sanford 1969; Poole et al. 1970).

The Hudson Platform also experienced extensive reef development, with bank and barrier reef deposits (Attawapiskat Formation) completely encircling Hudson Bay Basin and fringing the northern side of the Moose River Basin (Figure 6, Map 2582, map case; Sanford 1987; Sanford and Grant 1990).

Maximum flooding of the craton during the early Ludlow(?) was abruptly terminated by widespread uplift and block faulting along the major arch systems bordering and intersecting the St. Lawrence and Hudson platforms (Sanford et al. 1985; Sanford 1987). This epeirogenic event was widespread, including much of North America and Greenland, and may have occurred in response to continental collision during an early phase of the Acadian Orogeny (the Salinic Disturbance?; Rodgers 1970). Vertical movement of the Severn, Fraserdale, Cape Henrietta Maria, Frontenac and Algonquin arches (*see* Figure 20.2) appears *to have modified the pattern of subsequent marine* transgressions and deposition in the area and marked the end of Sequence 5.

Lithologic details of Sequence 5 units are presented in Table 20.5. The present regional distribution of these units are shown on Maps 2541 to 2544 (map case). The distribution of the sequence as a whole is shown on Maps 2575 to 2578 (map case). The maximum pre-erosional and present distribution of the sequence is shown on Figure 6, Map 2582 (map case). The stratigraphy of Sequence 5 in the Appalachian Basin (Niagara Peninsula), Michigan Basin (Bruce Peninsula), Lake Timiskaming outlier and the Hudson Bay and Moose River basins is shown in Figure 20.24 and discussed in the following section[1].

## Geology

### SOUTHERN ONTARIO (NIAGARA PENINSULA AREA)

The lower units of Sequence 5, in the Niagara Peninsula area of southern Ontario (i.e., Thorold through to Decew formations; *see* Figure 20.24), are assigned to the Clinton Group (Bolton 1957). Most of these units pinch out or are truncated by unconformities to the west and north. Some workers (e.g., Liberty and Bolton 1971; Kilgour 1972; Brett et al. 1990) have considered the Dyer Bay to St. Edmund formations (of Sequence 4), or younger (?) strata, such as the lower part of the Amabel Formation, to be Clinton Group time equivalents. However, supportive biostratigraphic evidence, especially for the Amabel strata, has been complicated by extensive dolomitization. The only unit, outside of southwestern Ontario, which is indisputably considered a

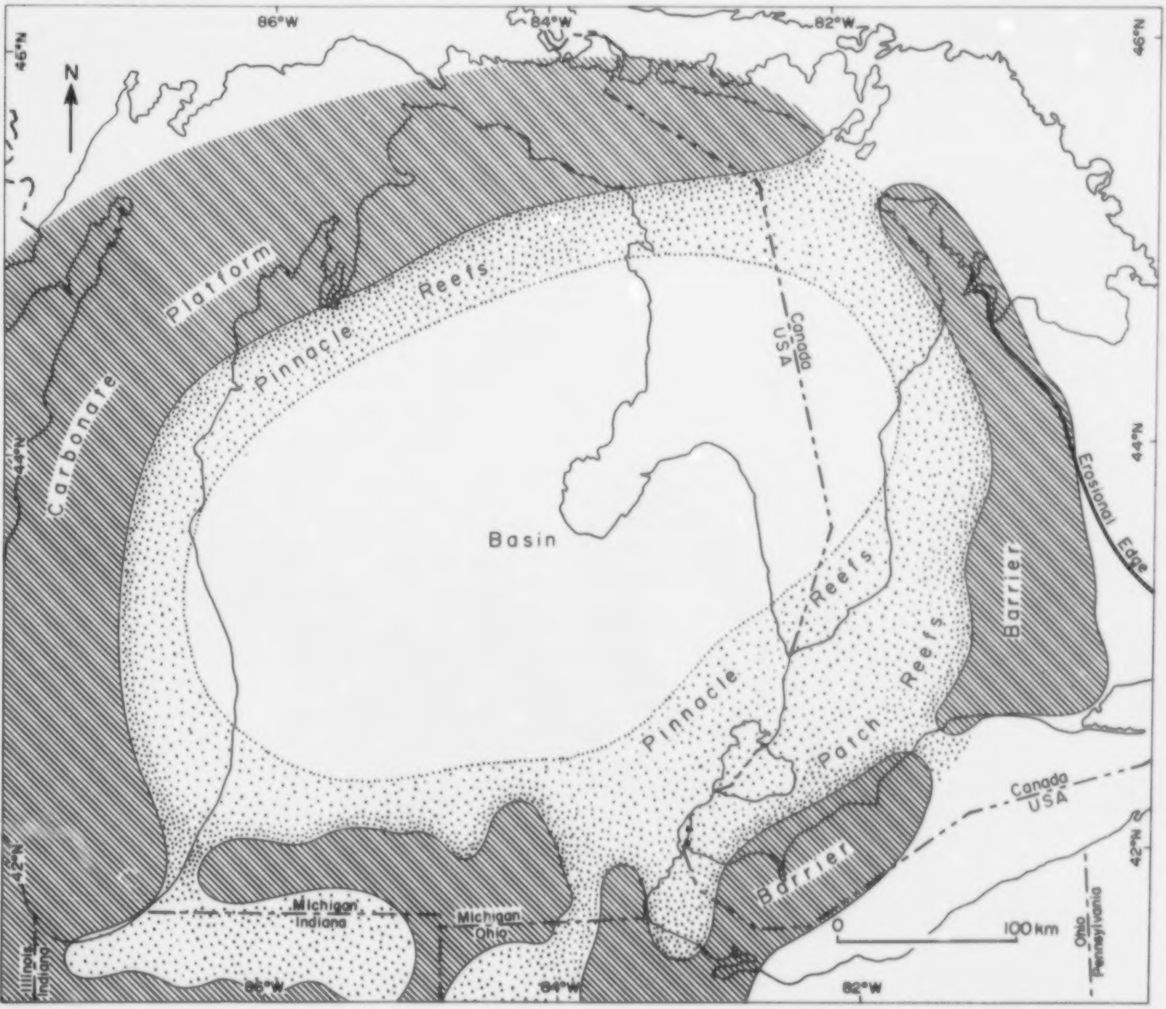
86°W
84°W
82°W
46°N
44°N
42°N
N
Carbonate
Platform
Pinnacle
Reefs
Canada
USA
Basin
Reefs
Pinnacle
Patch
Reefs
Barrier
Erosional Edge
Barrier
Canada
USA
Michigan
Indiana
Michigan
Ohio
Illinois
Indiana
Ohio
Pennsylvania
0
100 km

**Figure 20.23.** Distribution of upper Middle Silurian (upper Sequence 5) carbonate facies in and around the Michigan Basin (*modified from* Gill 1985; Sanford 1969; Bailey and Cochrane 1990).

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[1] *Note: the usage of rock stratigraphic terms Lower, Middle and Upper Silurian is discussed in the Silurian Terminology section in Sequence 4.*

**Table 20.5.** Middle Silurian (Llandovery to Ludlow) units in Ontario—Depositional Sequence 5 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Southern Ontario** | | | | | | |
| Guelph Formation | Southwestern Ontario (Niagara and Bruce peninsulas + subsurface) | Wenlock to Ludlow | Reef and interreef environments | 4 to 100 m | Buff to brown, fine- to medium-crystalline, saccharoidal dolostone; medium to thick bedded; richly fossiliferous (corals, stromatoporoids, bivalves, cephalopods).<br>Eramosa Mb. (Bruce Peninsula): thin- to medium-bedded, light tan to dark brown, fine- to medium-crystalline, sparsely to moderately fossiliferous, bituminous dolostone, locally laminated, cherty.<br>Lower contact with Eramosa Mb. of Lockport Fm. (Niagara Peninsula) and Amabel Fm. (Bruce Peninsula) is gradational. | Liberty and Bolton 1971; Armstrong and Goodman 1990; Sanford 1969 |
| Lockport Formation | Niagara Peninsula and subsurface of southwest Ontario | Wenlock to Ludlow | Range from shallow, moderate to high energy (crinoidal shoal + bioherm) to low energy restricted(?) subtidal environments | Eramosa Mb.: up to 20 m<br>Goat Island Mb.: up to 16 m<br>Gasport Mb.: up to 10 m | Consists of 3 members:<br>Eramosa Mb. (upper member): tan to black, fine-crystalline, bituminous, argillaceous dolostone; thin to medium bedded; locally laminated, cherty; gradational contact with underlying Goat Island Mb.<br>Goat Island Mb. (middle member): dark to light grey, very fine- to microcrystalline (sublithographic), locally shaly (Vinemount shale beds), or cherty (Ancaster chert beds) dolostone; thick and irregular bedding. Contact with Gasport Mb. is gradational to sharp.<br>Gasport Mb. (lower member): blue-grey (fresh surface), fine- to coarse-grained, massive-bedded dolomitic and argillaceous limestone; grainstone with cross-stratification; local biohermal developments.<br>Lower contact with Decew Fm. is sharp and erosive. | Bolton 1957; Crowley 1973; Telford 1978; Brett et al. 1990, 1991; Liberty and Bolton 1971 |
| Decew Formation (Clinton Group) | Niagara Peninsula | Wenlock | Shallow, restricted environment, offshore from crinoidal sand shoals of Gasport Mb. (Lockport Fm.) | up to 4 m | Dark grey, very fine-crystalline, argillaceous to arenaceous dolostone; shale content increases towards base of unit.<br>Soft-sediment deformation features; sparsely fossiliferous.<br>Lower contact is locally sharp to gradational with shale of Rochester Fm. | Bolton 1957, 1964; Crowley 1973 |

**Table 20.5.** Middle Silurian (Llandovery to Ludlow) units in Ontario—Depositional Sequence 5 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Rochester Formation (Clinton Group) | Niagara Peninsula and subsurface of southwestern Ontario | Wenlock | Storm-influenced, moderately shallow, muddy shelf | up to 24 m | Dark grey to black, calcareous shale and siltstone with abundant carbonate (calcarenitic and calcisiltic) interbeds.<br>Richly fossiliferous (brachiopods, bryozoans, corals, echinoderms, trilobites); bioturbated; tempestites, guttercasts, flute and groove molds, parallel, wavy- and cross-lamination, cut-and-fill structures.<br>Upper (Burleigh Hill) member.: moderately fossiliferous, blue-grey, calcareous shale with calcisiltite interbeds.<br>Lower (Lewiston) member.: fossiliferous, brown-grey shale with fossiliferous limestone interbeds.<br>Lower contact is sharp yet conformable with Irondequoit Fm. | Brett 1983;<br>Bolton 1957;<br>Sanford 1969 |
| Irondequoit Formation (Clinton Group) | Niagara Peninsula and subsurface of south-western Ontario | Wenlock | Shallow (near wave-base), high energy, crinoidal sand shoal environment | up to 3 m | Pinkish to light grey, porous, massive-bedded, crinoidal limestone; locally dolomitic; Reynales-like pebbles or intraclasts in basal part of unit; local development of bryozoan bioherms at top.<br>Lower contact is sharp, "welded", disconformable contact with Reynales Fm. | Bolton 1957; Mosher 1972; Liberty, Feenstra and Telford 1976a, 1976b; Liberty, Bond and Telford 1976a;<br>Brett et al. 1991 |
| Reynales Formation (Clinton Group) | Niagara Peninsula and subsurface of south-western Ontario | Llandovery | Subtidal to basinal environment | up to 5 m | Dark grey-brown, argillaceous dolostone and dolomitic limestone with greenish silty and/or shaly interbeds and partings; pyrite- and glauconite-rich and more argillaceous in lower part of unit. Sparsely to moderately fossiliferous (pentamerid brachiopods).<br>Lower contact conformable(?) with Neahga or Thorold Fms. | Bolton 1957;<br>Sanford 1969;<br>Brett et al. 1990 |
| Neahga Formation (Clinton Group) | Niagara Peninsula | Llandovery | Transgressive shallow, muddy shelf | up to 2 m | Olive green, fissile shale with minor grey limestone interbeds; relatively unfossiliferous; shale intraclasts.<br>Lower contact is well defined and conformable(?) with Thorold Fm. | Bolton 1957;<br>Fisher 1953;<br>Brett and Calkin 1987 |
| Thorold Formation (Clinton Group) | Niagara Peninsula and east-central Lake Erie area | Llandovery | Very shallow, tidal sand shoal or sandy intertidal flat | up to 6.5 m (3 m avg.) | Grey-green to white, fine- to coarse-grained, quartzose sandstone; green-grey glauconitic shale and siltstone partings may be present. Internally channelled; marine trace fossils; trough, ripple and hummocky cross-stratification; flat laminations; herringbone cross-bedding; shale intraclasts.<br>Lower contact is sharp, erosional and disconformable with the Grimsby Fm. (Sequence 4). | Sanford 1969;<br>Duke et al. 1991;<br>Pemberton 1987;<br>Martini 1971;<br>Duke 1991 |

**Table 20.5.** Middle Silurian (Llandovery to Ludlow) units in Ontario—Depositional Sequence 5 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Amabel Formation | Manitoulin Island and Bruce Peninsula to Burlington + subsurface of southwestern Ontario | Wenlock | Shallow, high energy shoal to deeper, basinal(?) environments | Wiarton/Colpoy Bay Mb.:<br>up to 25 m<br>Lions Head Mb.:<br>up to 13 m | Consists of 2 members:<br>Wiarton/Colpoy Bay Mb. (upper member): light grey, blue-grey mottled, medium- to coarse-crystalline dolostone; crinoidal and locally biohermal; thick to massive bedded; dolomitization obscures fossil content (echinoderms, corals, brachiopods); commonly cross-stratified.<br>Lions Head Mb. (lower member): tan to light grey-brown, fine-crystalline dolostone; thin bedded; sparsely fossiliferous; burrowed; chert nodules.<br>Lower contact with Fossil Hill Fm. sharp and disconformable. | Bolton 1957;<br>Liberty and Bolton 1971;<br>Armstrong 1987, 1988, 1989;<br>Armstrong and Goodman 1990 |
| Fossil Hill Formation | Bruce Peninsula, Manitoulin Island + subsurface of southwestern Ontario | Llandovery | Shoaling: low energy, below wave base, to high energy, above wave base | up to 24 m<br>(<3 m avg.) | Buff to grey-brown, thin- to medium-bedded, fine- to coarse-crystalline dolostone; richly fossiliferous (brachipods, corals, stromatoporoids); coral biostromes; silicified fossils; locally unfossiliferous micritic beds (False Mindemoya Beds) near base of unit.<br>Lower contact with St. Edmund and older fms. (Sequence 4) is sharp and unconformable (may be conformable with St. Edmund Fm. on Manitoulin Island). | Sanford 1969;<br>Johnson 1979;<br>Johnson and Telford 1981a, 1981b, 1985a-c,e,f;<br>Armstrong and Goodman 1990 |
| **Northern Ontario** | | | | | | |
| Attawapiskat Formation | Hudson Bay and Moose River basins | Wenlock to Ludlow | Reef and interreef environments | up to 62 m | Tan to brown biohermal limestone and dolostone; bioclastic flank and interbiohermal (lime mudstone and dolostone) deposits. Moderately to richly fossiliferous.<br>Lower contact conformable with Ekwan River Fm. | Sanford 1987;<br>Sanford et al. 1968;<br>Norris 1986;<br>Chow and Stearn 1989 |
| Ekwan River Formation | Hudson Bay and Moose River basins | Llandovery to Ludlow | Platform | up to 235 m | Tan, cream and brown, fine- to medium-grained limestone and dolostone; locally bioclastic, biostromal or bituminous; chert nodules, richly fossiliferous.<br>Lower contact with Severn River Fm. is conformable(?). | Sanford 1987;<br>Norris 1986 |
| Thornloe Formation | Lake Timiskaming outlier | Wenlock | Undetermined | ~70 m | Buff to tan, very fine-crystalline dolostones; massive to thick bedded; some fossils (stromatolites); chert nodules.<br>Lower contact disconformable with Earlton Fm. | Russell 1984 |
| Earlton Formation | Lake Timiskaming outlier | Llandovery | Undetermined | ~62 m | Grey, very fine- to microcrystalline (sublithographic) limestone and fine-laminated dolostone with bioclastic and oolitic interbeds and shaly partings; thin bedded; fossiliferous. Fossils include corals, gastropods and brachipods.<br>Lower contact with Evanturel Creek Fm. (Sequence 4) is unconformable. | Russell 1984;<br>Bolton and Copeland 1972 |

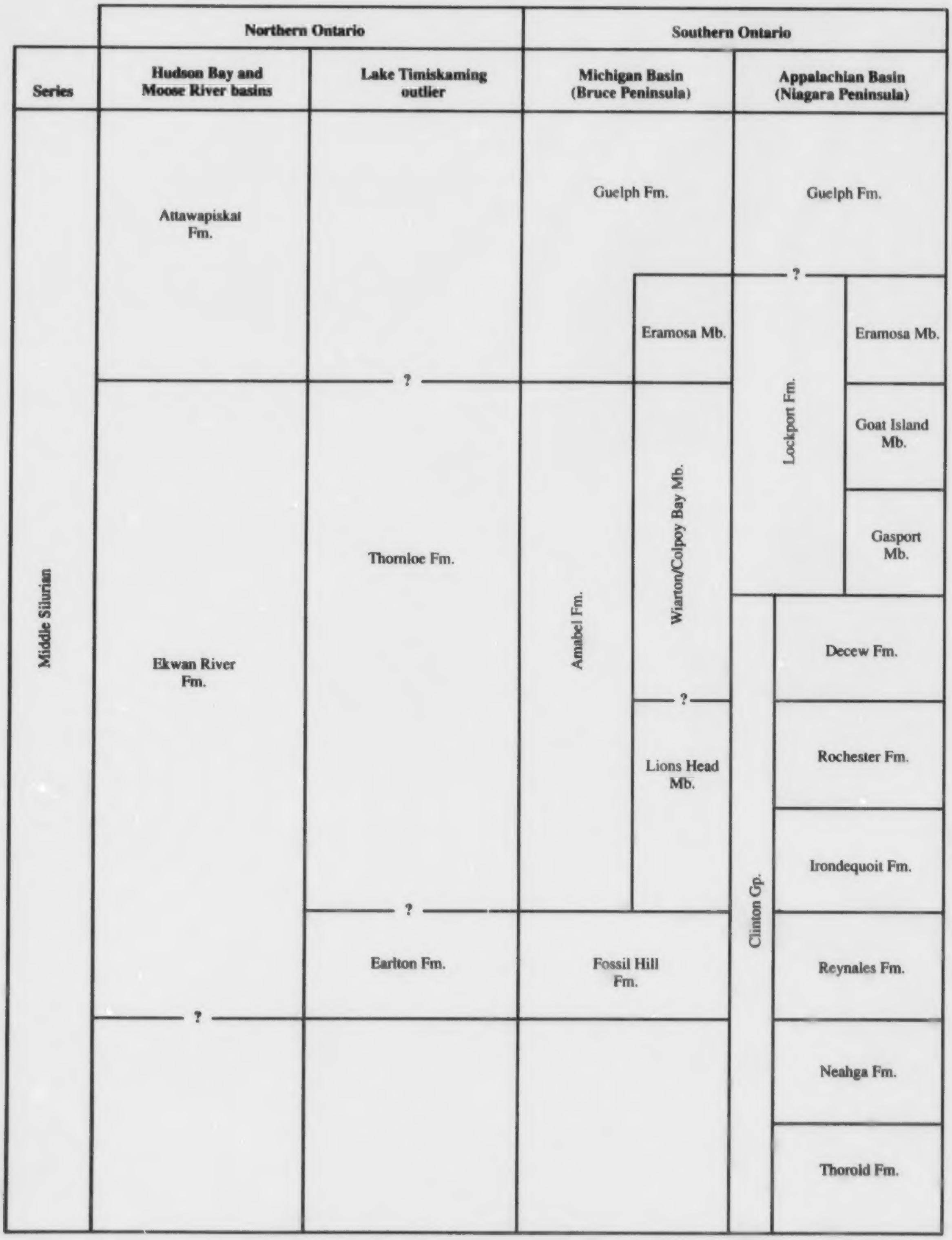
Northern Ontario
Southern Ontario
Series
Hudson Bay and Moose River basins
Lake Timiskaming outlier
Michigan Basin (Bruce Peninsula)
Appalachian Basin (Niagara Peninsula)
Middle Silurian
Attawapiskat Fm.
Ekwan River Fm.
?
Thornloe Fm.
Earlton Fm.
?
Guelph Fm.
Eramosa Mb.
Amabel Fm.
Wiarton/Colpoy Bay Mb.
Lions Head Mb.
Fossil Hill Fm.
Guelph Fm.
?
Lockport Fm.
Eramosa Mb.
Goat Island Mb.
Gasport Mb.
Clinton Gp.
Decew Fm.
Rochester Fm.
Irondequoit Fm.
Reynales Fm.
Neahga Fm.
Thorold Fm.

**Figure 20.24.** Stratigraphic units of Sequence 5 (? indicates uncertain age).

Clinton Group equivalent is the Fossil Hill Formation. Due to these uncertainties, the term Clinton Group is herein not extended outside of the Niagara Peninsula area (*see* Figure 20.24; Chart D, Map 2582, map case). The relationship of Clinton Group units of the Niagara Peninsula area to Sequence 5 strata elsewhere in southern Ontario is discussed at the end of this section.

The shallow marine sandstones of the Thorold Formation form the base of Sequence 5 in the Niagara Peninsula and east-central Lake Erie areas of southern Ontario (*see* Figure 20.24; Sanford 1969). The Thorold Formation is exposed along the base of the Niagara Escarpment from the Niagara Gorge to just west of Hamilton, where it grades laterally and vertically into the Reynales Formation (Sanford 1969). This has led to the suggestion that the Thorold Formation may be a nearshore facies of the Reynales Formation.

The Thorold Formation is sheet-like in character and averages about 3 m thick. The unit is complexly channelled and contains marine trace fossils, particularly in the Hamilton area (Pemberton 1987; Duke et al., in press). Sedimentary structures include trough cross-bedding, flat-lamination and hummocky cross-stratification. Depositional environments proposed for the unit include longshore bar and beach (Martini 1971) and shallow tidal sand shoal (Duke et al., in press).

Some researchers (e.g., Caley 1940; Martini 1971; Kilgour 1972; Duke 1987; Duke et al., in press), believing the Thorold Formation to be lithologically and sedimentologically related to the Lower Silurian clastic units of Sequence 4, group the Thorold Formation with these underlying clastic rocks, thereby forming the Medina Group (*see* Sequence 4). Textural evidence, however, suggests that the Thorold Formation contains reworked Grimsby Formation sediments (Fisher 1954; Bolton 1957; Sanford 1969; Brett and Calkin 1987) and indicates at least a brief break in sedimentation between these 2 units. The lower contact of the Thorold Formation with the Grimsby Formation is also discussed in the Sequence 4 description. The upper contact with the Neahga Formation shale or (farther to the west) the Reynales Formation dolostone is relatively sharp and well defined (Bolton 1957). Some workers (e.g., Brett et al. 1990) believe that the Thorold, Neahga and Reynales formations are separated by disconformities.

The overlying Neahga Formation, a fissile shale with minor limestone interbeds, is restricted in Ontario to the Niagara Peninsula area (*see* Figures 20.24 and 20.25). This unit thins westward from about 2 m at Niagara Falls to its pinch-out near Grimsby Beach (Bolton 1957). It appears to represent a short transgressive pulse of shallow marine clastic sediments deposited below wave base and confined to the Appalachian Basin (Fisher 1954; Brett and Calkin 1987). The Neahga Formation shales are seen at only a few localities in Ontario; the Sir Adam Beck Generating Station access road in the Niagara Gorge shows the best exposure (*see* Figure 20.25; Bolton 1964; Telford 1978).

Overlying the Neahga Formation shales and the Thorold Formations sandstones are the argillaceous dolostones and dolomitic limestones of the Reynales Formation. This unit occurs in the subsurface of southwestern Ontario and the Niagara Peninsula area (Sanford 1969) and outcrops along the Niagara Escarpment (*see* Figure 20.1) from the Niagara Gorge to near Georgetown (Bolton 1957). Overall, the Reynales Formation is sparsely fossiliferous; however, the lower beds commonly contain pentamerid brachiopods (Bolton 1957). The basal bed of this formation is more argillaceous, pyritic and glauconitic, suggesting deposition under reducing conditions. The Reynales Formation is considered to be equivalent to and a basinal facies of the Fossil Hill Formation (Sanford 1969).

Recent detailed work on the Niagara Peninsula (Brett et al. 1990) suggests that the Reynales Formation of New York state is truncated by an unconformity just west of the Niagara Gorge and that the Reynales Formation of Ontario actually consists of 2 younger units. The lower unit, containing pentamerid brachiopods, was termed the Merritton Member of the Reynales Formation by Kilgour (1963, 1972) and the Merritton Formation by Brett et al. (1990). The upper unit has been termed the Rockway Member of the Irondequoit Formation (Kilgour 1963). Brett et al. (1990) and Kilgour (1963) suggested that the pentamerid brachiopod-bearing Merritton Member (or Formation) directly correlates with the Fossil Hill Formation and that the Rockway Member correlates with the Lions Head Member of the Amabel Formation. The terms Merritton Member (or Formation) and Rockway Member are not currently used by the Ontario Geological Survey; adoption of this revised stratigraphy and these correlations is pending better constrained biostratigraphic evidence.

In southern Ontario, a depositional hiatus of unknown duration occurs at the top of the Reynales and Fossil Hill formations (Chart D, Map 2582, map case). After this hiatus in the Niagara Peninsula area (and in part to the southwest), sedimentation continued with deposition of (in ascending order) the Irondequoit, Rochester, Decew, Lockport and Guelph formations (*see* Figures 20.24 and 20.25; Bolton 1957). Elsewhere in southern Ontario, and at least as far north as the Lake Timiskaming outlier, sedimentation continued with the deposition of a thick succession of carbonate sediments, represented by the Lockport, Amabel, Guelph and Thornloe formations (Chart D, Map 2582, map case; *see* Figure 20.24).

Overlying the Reynales Formation in the Niagara Peninsula area is the relatively thin Irondequoit Formation, a porous, pink crinoidal limestone (Bolton 1957). This formation commonly occurs as 1 or 2 massive, erosion-resistant beds, which outcrop as a subescarpment below the main brow of the Niagara Escarpment (Hewitt 1971). This unit also extends in the subsurface to just east of London (Sanford 1969). The Irondequoit Formation is separated from the underlying Reynales Formation by an unusual "welded contact", a term used to describe a lithologic change without a bedding plane separation (Mosher 1972; Liberty, Feenstra and Telford 1976a, 1976b; Liberty, Bond and Telford 1976a), above which pebbles, or intraclasts, of Reynales-like material are present within the basal Irondequoit Formation. The Irondequoit Formation thins from east

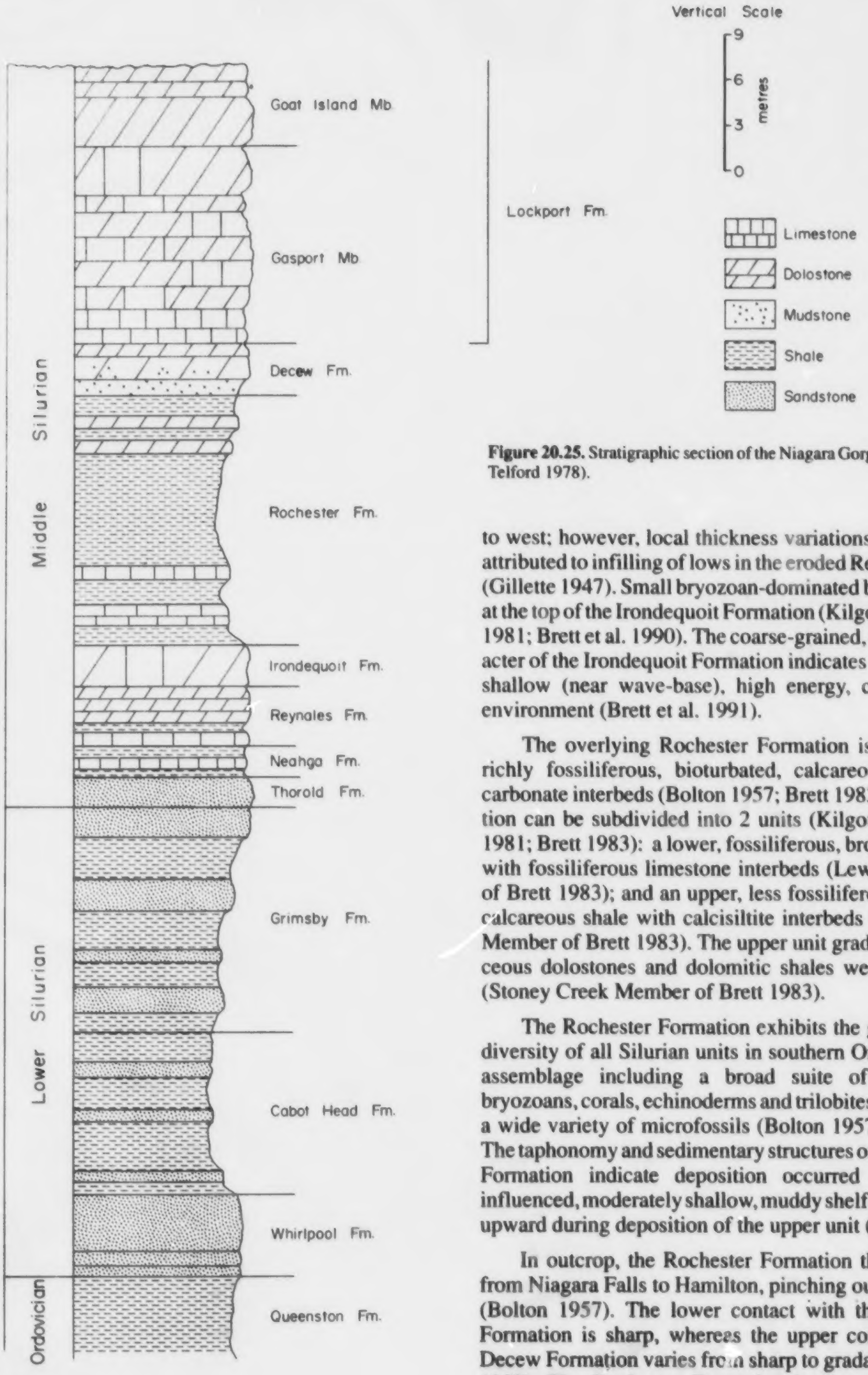
Vertical Scale
9
6
3
0
metres
Limestone
Dolostone
Mudstone
Shale
Sandstone
Goat Island Mb.
Gasport Mb.
Lockport Fm.
Decew Fm.
Rochester Fm.
Irondequoit Fm.
Reynales Fm.
Neahga Fm.
Thorold Fm.
Grimsby Fm.
Cabot Head Fm.
Whirlpool Fm.
Queenston Fm.
Middle Silurian
Lower Silurian
Ordovician

**Figure 20.25.** Stratigraphic section of the Niagara Gorge (*modified from* Telford 1978).

to west; however, local thickness variations occur and are attributed to infilling of lows in the eroded Reynales surface (Gillette 1947). Small bryozoan-dominated bioherms occur at the top of the Irondequoit Formation (Kilgour and Liberty 1981; Brett et al. 1990). The coarse-grained, crinoidal character of the Irondequoit Formation indicates deposition in a shallow (near wave-base), high energy, crinoidal shoal environment (Brett et al. 1991).

The overlying Rochester Formation is a sparsely to richly fossiliferous, bioturbated, calcareous shale with carbonate interbeds (Bolton 1957; Brett 1983). This formation can be subdivided into 2 units (Kilgour and Liberty 1981; Brett 1983): a lower, fossiliferous, brown-grey shale with fossiliferous limestone interbeds (Lewiston Member of Brett 1983); and an upper, less fossiliferous, blue-grey, calcareous shale with calcisiltite interbeds (Burleigh Hill Member of Brett 1983). The upper unit grades into argillaceous dolostones and dolomitic shales west of Grimsby (Stoney Creek Member of Brett 1983).

The Rochester Formation exhibits the greatest faunal diversity of all Silurian units in southern Ontario, with an assemblage including a broad suite of brachiopods, bryozoans, corals, echinoderms and trilobites, in addition to a wide variety of microfossils (Bolton 1957; Brett 1983). The taphonomy and sedimentary structures of the Rochester Formation indicate deposition occurred on a storm-influenced, moderately shallow, muddy shelf that shallowed upward during deposition of the upper unit (Brett 1983).

In outcrop, the Rochester Formation thins westward from Niagara Falls to Hamilton, pinching out near Dundas (Bolton 1957). The lower contact with the Irondequoit Formation is sharp, whereas the upper contact with the Decew Formation varies from sharp to gradational (Bolton 1957). The Rochester Formation is also present in the subsurface of southwestern Ontario (Sanford 1969), where as a distinct unit, it is an important stratigraphic marker on geophysical logs (Beards 1967; Bailey and Cochrane 1986).

The Rochester Formation is conformably overlain by the Decew Formation (*see* Figures 20.24 and 20.25), a thin, argillaceous dolostone. This unit is geographically restricted to the northern rim of the Appalachian Basin and outcrops along the Niagara Escarpment as far west as Hamilton. Locally, the Decew Formation displays spectacular soft-sediment deformation or "enterolithic" features (Kilgour and Liberty 1981). The unit is sparsely fossiliferous and shale content increases below Lake Erie, where it becomes indistinguishable from the underlying Rochester Formation (Sanford 1969).

In the Niagara Peninsula area, the upper contact of the Decew Formation with the Gasport Member of the Lockport Formation (*see* Figures 20.24 and 20.25) is commonly sharp and erosive (Bolton 1957; Crowley 1973) and the variation in thickness of the Decew Formation has been attributed to post-depositional erosion (Bolton 1957). Crowley (1973) suggested that the Decew muddy carbonate sediments were deposited in a relatively shallow, restricted environment, just offshore from crinoidal sand shoals of the overlying Gasport Member of the Lockport Formation, which subsequently prograded out over the Decew Formation.

The carbonate rocks of the Lockport Formation form the brow and main cliffs of the Niagara Escarpment on the Niagara Peninsula (Bolton 1957; Hewitt 1971) and extend in the subsurface under much of southwestern Ontario (Sanford 1969). The Lockport Formation consists of a complex group of lithofacies, including coarse grainstones, bioherms and argillaceous, bituminous dolostones (Bolton 1957; Crowley 1973; Brett et al. 1991), which have been grouped into 3 members: these are in ascending order, the Gasport, Goat Island and Eramosa members. Two informal units have been identified near the top of the Goat Island Member: the Ancaster chert beds (Bolton 1957) and the Vinemount shale beds (Telford 1978). Depositional environments interpreted for the Lockport Formation range from shallow high energy to partly restricted subtidal environments (Crowley 1973; Brett et al. 1990).

The lowest member of the Lockport Formation, the Gasport Member, consists of massive, blue-grey, crinoidal dolomitic limestones with local bioherms (Bolton 1957; Crowley 1973). Some bioherms exhibit evidence of brief subaerial exposure (Crowley 1973). The basal contact of the Gasport Member with the Decew Formation is sharp and erosive with a locally developed intraclastic conglomerate.

The overlying Goat Island Member is a thinner bedded, finer crystalline, grey dolostone which is locally cherty or shaly (Bolton 1957). The contact between the Goat Island and Gasport members varies from transitional to sharp and Gasport-like lithofacies (e.g., grainstone beds) are reported within the Goat Island interval (e.g., Brett et al. 1990). The Vinemount shale beds and Ancaster chert beds locally cap the Goat Island Member. Due to the generally poor exposure of both of these units, their lateral distribution is not well known (Liberty, Feenstra and Telford 1976a, 1976b; Liberty, Bond and Telford 1976a).

The uppermost member of the Lockport Formation is the Eramosa Member, a bituminous, argillaceous, thin-bedded dolostone (Bolton 1957). The Eramosa Member has gradational contacts with the underlying Goat Island Member and the overlying Guelph Formation. Northwest of the Niagara Peninsula, bituminous strata at this stratigraphic position have been assigned to the Eramosa Member of the Amabel Formation (Bolton 1957; Liberty and Bolton 1971); although this assignment is in dispute (e.g., Armstrong and Goodman 1990; Sanford 1969). The bituminous nature of the Eramosa Member suggests deposition in at least a partly restricted marine environment. Northwest of the Niagara Peninsula, this unit was more clearly deposited in restricted interbiohermal (or inter-reefal) positions (Liberty and Bolton 1971; Armstrong and Goodman 1990).

The Guelph Formation, which overlies the Eramosa Member, is poorly exposed in the Niagara Peninsula area (Liberty, Feenstra and Telford 1976a, 1976b). To the northwest, in its type area at Guelph, biostromal (corals, stromatoporoids, gastropods, bivalves), thicker dolostone beds of this formation conformably overlie bituminous, argillaceous, less fossiliferous dolostones of the Eramosa Member (Telford 1976b). The upper contact of the Guelph Formation with the Salina Formation (of Sequence 6) is not exposed in the Niagara Peninsula area. The Guelph Formation is better exposed to the northwest and is discussed more fully in the next section.

As previously mentioned, there is some uncertainty concerning the correlation of Sequence 5 units both within the Niagara Peninsula area (including part of the subsurface of southwestern Ontario) and with Sequence 5 strata elsewhere in southern Ontario. Generally, the Niagara Peninsula succession is more clastic rich, reflecting its location on the northwestern flank of the Appalachian Basin and its proximity to the Appalachian Orogen. Rapid lateral facies changes and depositional pinch-outs across the peninsula indicate that the Algonquin Arch (*see* Figure 20.2) was a positive topographic feature during this time, forming the northwestern margin of the Appalachian Basin. Also, as relative sea level fluctuated (tectonically induced?), erosional disconformities developed on topographic highs such as the arch. Thus, a complex succession resulted where generally thin, condensed, sparsely fossiliferous units either change character laterally, pinch out, or are cut by erosional disconformities. The relatively short period during which this succession was deposited makes it difficult to date the units or the disconformities and, thus, to correlate them either across the peninsula or to equivalent successions elsewhere.

The contacts between many of these units have been variously interpreted to represent major depositional hiatuses or minor erosive events (e.g., storms). Recent workers (e.g., Brett et al. 1990) have recognized laterally equivalent facies belts which genetically link some units and identified disconformities which separate previously combined units (e.g., the Thorold, Neahga, Reynales "package") and, thus, have interpreted a number of small-scale depositional sequences (or subsequences) comprising Sequence 5 in this area. Dolomitization of approximately equivalent units (e.g., Amabel Formation)

**Figure 20.26.** Silicified brachiopods and corals in an outcrop of the Fossil Hill Formation on Manitoulin Island. (Photo: DA89–18–15)

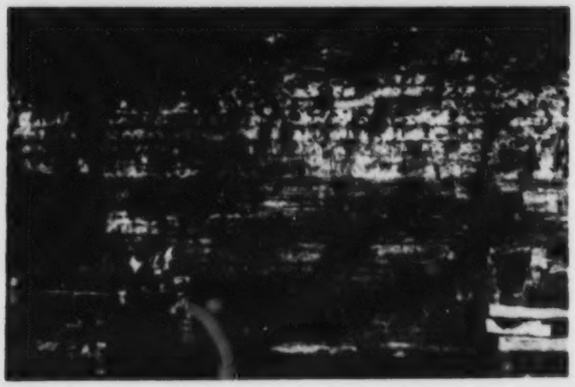

**Figure 20.27.** Vuggy, nonlaminated dolostone beds overlying 2 m of laminated dolostone beds, both of the Eramosa Member, Guelph Formation exposed in the Ross Quarry, northwest of Wiarton, Bruce Peninsula. (Photo: DA86–35)

further complicates correlation of Niagara Peninsula units outside this area.

## SOUTHERN ONTARIO (EXCLUDING THE NIAGARA PENINSULA AREA)

In southern Ontario, outside of the Niagara Peninsula and Lake Erie area, carbonate deposition predominated during Sequence 5 time (Figure 6 and Chart D, Map 2582, map case). Generally, during this time, deeper water carbonate sediments were deposited in the more rapidly subsiding basin centres, whereas various shelf and platformal facies were deposited in the shallower margins of the basins (e.g., the Algonquin Arch). This basin-centred facies differentiation, which appears to have affected the distribution patterns of the Fossil Hill and Reynales formations (Sanford 1969), became more pronounced later, when pinnacle and patch reefs (Guelph Formation) developed along "hinge lines" separating basins and arches. The distribution patterns of these later facies in the Michigan Basin is shown in Figure 20.23. Recent detailed mapping of these carbonate strata in outcrop along the northern and eastern margins of the Michigan Basin (Johnson and Telford 1985a–f; Wolf 1986; Armstrong and Goodman 1990; Armstrong and Dubord 1992; Armstrong, in prep. a, in prep. b, in prep. c) has shown local facies relationships to be complex both laterally and vertically.

The oldest Sequence 5 deposits outside of the Niagara Peninsula area consist of the fossiliferous dolostones of the Fossil Hill Formation. This unit outcrops extensively on Manitoulin Island (Johnson and Telford 1985a–f) and Cockburn Island (Wolf 1986), and is progressively less well exposed southward (Bolton 1957; Liberty and Bolton 1971; Bond et al. 1976; Armstrong 1987, 1988, 1989). The lateral transition of the Fossil Hill to the Reynales Formation is not well understood, partly because of the poor, intermittent exposure of these units along the outcrop belt between Burlington and Collingwood (Bond et al. 1976; Bond and Telford 1976; Liberty, Bond and Telford 1976a, 1976b; Telford 1976a). In the subsurface, the Fossil Hill Formation occurs in a broad belt trending southward from its outcrop area to central Lake Erie (Sanford 1969). To the southeast and west of this broad belt, into the Appalachian and Michigan basins, respectively, the Fossil Hill Formation grades laterally into the more basinal Reynales Formation (Sanford 1969).

The Fossil Hill Formation contains thick pentamerid brachiopod banks which commonly grade upward into coral and stromatoporoid biostromes. Evolution of pentamerid brachiopod shell morphology, identified within the Fossil Hill Formation (Johnson and Colville 1982), may be attributed to environmental stress related to fluctuations in relative sea level (Johnson 1981). The change from brachiopod banks, formed at or below wave base, to coral and stromatoporoid biostromes, formed above wave base, has also been attributed to shoaling (Johnson 1979).

The Fossil Hill Formation is also known for its chert content and well-preserved, silicified fauna (Figure 20.26; Johnson 1979). Locally, the Fossil Hill Formation contains up to 2 m of a lithofacies similar to the underlying St. Edmund Formation (*see* Sequence 4; Johnson and Telford 1981a, 1981b, 1985a–f). The incorrect identification of this bed, informally termed the "False Mindemoya", has led to the misplacement of the Fossil Hill–St. Edmund Formation contact on many earlier maps of Manitoulin Island (e.g., Liberty 1972a–f). This repetition of St. Edmund lithofacies within the Fossil Hill interval suggests a conformable relationship between these units.

On Manitoulin Island, the basal contact of the Fossil Hill Formation with the St. Edmund Formation is reported as sharp and conformable (Johnson and Telford 1985f). To the south, on the Bruce Peninsula, the basal contact of the Fossil Hill Formation is a regionally angular unconformity which progressively cuts the St. Edmund Formation and older units (*see* Sequence 4; Armstrong and Goodman 1990).

Overlying the Fossil Hill Formation are the thick, extensive dolostones of the Amabel Formation (Bolton 1957; Liberty and Bolton 1971). Although dolomitization has obscured much of the original depositional fabric and faunal remains in the Amabel Formation, crinoidal bank, basinal and biohermal lithofacies have been identified (Sanford 1969; Liberty and Bolton 1971).

The white weathering, thick-bedded Amabel dolostones form the main cliffs of the Niagara Escarpment from north of Burlington to Manitoulin and Cockburn Islands (*see* Figure 20.20; Map 2544, map case). On the Niagara Peninsula and to the southwest in the subsurface, the Amabel Formation is represented by the approximately equivalent Lockport Formation (Bolton 1957; Sanford 1969). The succession on the Niagara Peninsula from the Irondequoit Formation through to the Decew Formation (*see* Figure 20.24) may be time equivalent to the lower Amabel Formation (Chart D, Map 2582, map case).

On the Bruce Peninsula, 2 members are identified comprising the Amabel Formation: the Lions Head and the Wiarton/Colpoy Bay (Armstrong 1987, 1988, 1989; Armstrong and Goodman 1990). The lower Lions Head Member is a sparsely fossiliferous, fine-crystalline dolostone, which was probably deposited in a deeper offshore (basinal?) setting than the upper member. The overlying Wiarton/Colpoy Bay Member consists of massive beds of coarser crystalline, crinoidal and locally biohermal dolostone, which represent shallower, higher energy deposition (Armstrong and Goodman 1990). This member includes strata previously assigned to the Wiarton and Colpoy Bay members (e.g., Bolton 1957).

Surface mapping on the Bruce Peninsula indicates that the Wiarton/Colpoy Bay Member generally overlies the Lions Head Member (Liberty and Bolton 1971; Armstrong and Goodman 1990). However, drilling at Tobermory has shown that the Wiarton/Colpoy Bay Member grades laterally into the Lions Head, the latter constituting the entire Amabel interval at this locality (Sanford 1969; Armstrong 1989, in prep. a; Armstrong and Goodman 1990).

Along the outcrop belt between Collingwood and Georgetown, and in the subsurface, the Amabel Formation also consists of the above described rock types, but the cover of Quaternary sediments obscures their interrelationships (Bolton 1957; Liberty and Bolton 1971; Telford 1976a, 1976b; Bond et al. 1976; Bond and Telford 1976; Liberty, Bond and Telford 1976a, 1976b). On Manitoulin Island, the Amabel Formation is subdivided into several lithofacies (Johnson and Telford 1985a–f) with complex, poorly understood interrelationships.

Previously assigned to the Amabel Formation (Bolton 1957, Liberty and Bolton 1971) is the Eramosa Member, a generally bituminous, thin-bedded dolostone (Figure 20.27). Recent mapping of this member on the Bruce Peninsula suggests that these strata should be assigned to the overlying Guelph Formation (Armstrong 1987, 1988, 1989; Armstrong and Meadows 1988). This reassignment may also be applicable for approximately equivalent strata, in the Niagara Peninsula area, which are presently assigned to the Eramosa Member of the Lockport Formation. North of Guelph, Eramosa Member strata grade laterally and vertically into the Guelph Formation and are readily distinguished from the underlying Amabel Formation (Telford 1976b).

**Figure 20.28. Bioherms of the Guelph Formation exposed in roadcuts along Highway 6, northwest of Lion's Head, on the Bruce Peninsula. (Photo: DA89–34–30)**

The Eramosa Member occupies an interbiohermal position between Guelph Formation and possibly upper Amabel Formation bioherms (Liberty and Bolton 1971; Armstrong and Goodman 1990). Its organic-rich and laminated character is suggestive of a restricted shallow marine environment (lagoonal?). Locally, Amabel bioherms are directly overlain by Guelph bioherms and the Eramosa Member is absent (Liberty and Bolton 1971; Armstrong and Goodman 1990).

The Guelph Formation consists of reef and interreef deposits, characterized by tan, sugary, fossiliferous dolostone. Fossils within this unit include corals, stromatoporoids, bivalves (including the large *Megalomus canadensis*) and cephalopods (Liberty and Bolton 1971). Lower Guelph bioherms (Figure 20.28) are laterally equivalent with Eramosa Member strata. Upper Guelph strata conformably overlie the Eramosa Member (Sanford 1969; Armstrong and Goodman 1990).

Late Silurian biohermal development in southern Ontario culminated during the deposition of the Guelph Formation. Pinnacle and patch reefs proliferated on the periphery of the Michigan Basin and were, in turn, ringed by platform carbonate sediments (*see* Figure 20.23; Figure 6, Map 2582, map case; Sanford 1969; Gill 1985). In the basin centre, deeper water, basinal carbonate sediments were deposited. Guelph bioherms developed on topographic highs, which were themselves usually formed by underlying Amabel or Lockport biohermal mounds (Liberty and Bolton 1971). The detailed interrelationship between these carbonate deposits is complex, and not well understood.

A model recently proposed by Sanford et al. (1985) suggested that episodic tectonism through the Phanerozoic created fault-bounded blocks (Figure 20.29), which at times controlled local depositional patterns. Sanford et al. (1985) also suggested that Guelph pinnacle reefs were localized on the upthrown edges of these blocks (*see also* Summary and Discussion).

The upper contact of the Guelph Formation with the Salina Formation (of Sequence 6) is nowhere exposed in Ontario. In the subsurface of southwestern Ontario and adjacent areas of Michigan, the relationship between these 2 units, investigated in the course of hydrocarbon exploration, is complex and continues to be the subject of controversy (e.g., Mesolella et al. 1974; Gill 1977, 1985; Shaver et al. 1978; Sears and Lucia 1979; Droste and Shaver 1985). Guelph Formation pinnacle reefs (ranging from 90 to 180 m tall) are, for the most part, encased in evaporites, evaporitic carbonates and shales of the Salina Formation. Early workers tended to believe that the pinnacle reefs and the interreefal Guelph strata were separated from the Salina Formation by an unconformity. More recent consensus suggests that at least 1 unconformity exists within a typical Guelph pinnacle reef and that above this, reefal carbonates (assigned to the Guelph Formation) were deposited contemporaneously with encasing evaporites (of the Salina Formation). Some workers (e.g., Smith 1990) suggested multiple periods of subaerial exposure and karst development occurred during pinnacle reef growth. Thinning of lower Salina Formation units over the Algonquin Arch (Carter 1987; *see* Figure 20.32) and the development of very shallow to terrestrial facies in this unit in this area (Haynes et al. 1989), are consistent with the development of an unconformity between the Guelph and Salina formations over the Algonquin Arch (Sanford 1969; Chart D, Map 2582, map case). Thus, although the contact which marks the upper boundary of Sequence 5 in southern Ontario is not an unequivocal unconformity, it does, however, mark a fundamental change in sedimentation.

## NORTHERN ONTARIO

The oldest Sequence 5 unit in the Lake Timiskaming outlier, the Earlton Formation (*see* Figure 20.24; Chart D,

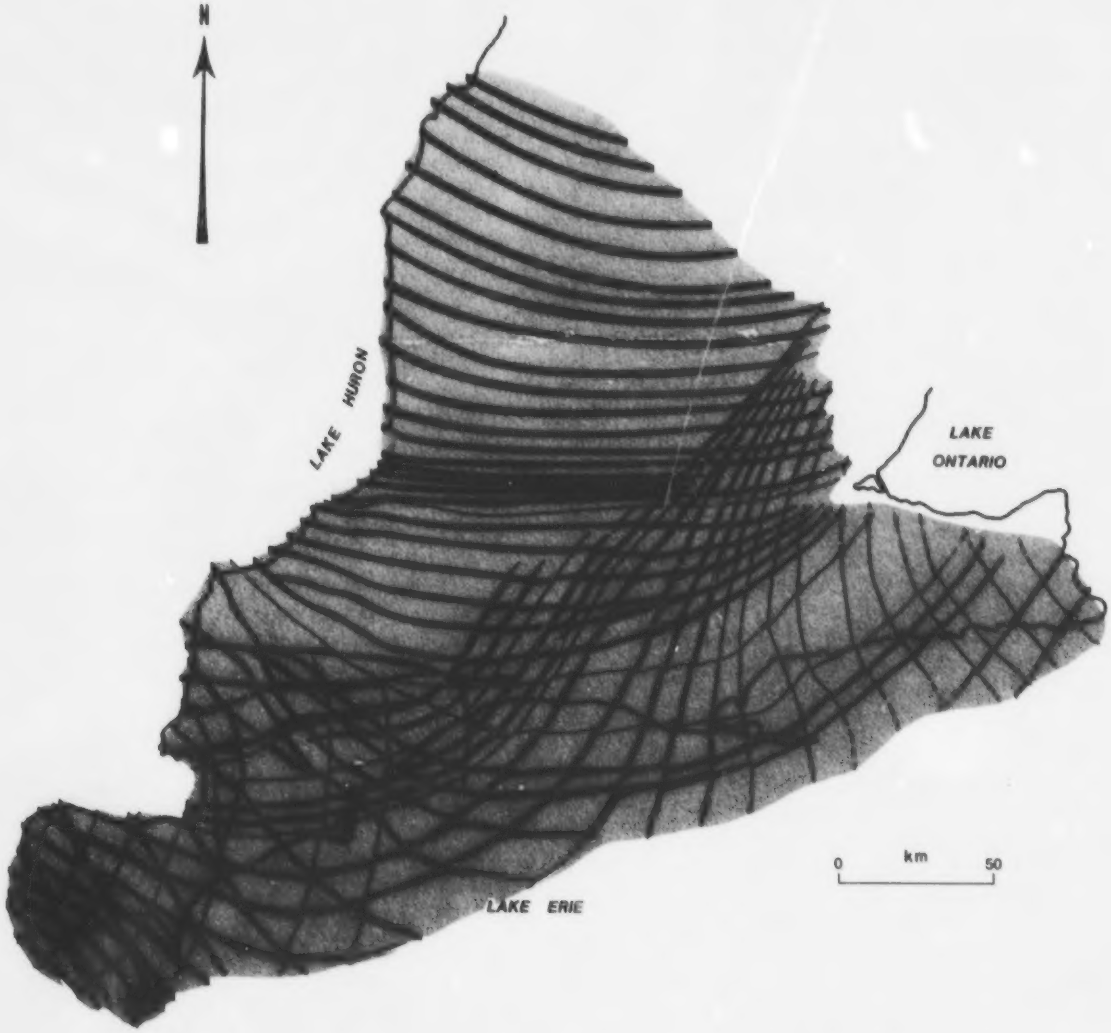

**Figure 20.29.** Conceptual fracture-framework model for southwestern Ontario (*modified from* Sanford et al. 1985).

Map 2582, map case), consists of very fine-grained limestone and very fine-crystalline dolostone with bioclastic and oolitic interbeds. Fauna include corals, gastropods and pentamerid brachiopods (Russell 1984), the latter suggesting correlation with the Fossil Hill Formation of southern Ontario. The basal part of the overlying Thornloe Formation also contains pentamerid brachiopods (Bolton and Copeland 1972; Russell 1984).

Within the Lake Timiskaming outlier, massive, escarpment-forming dolostones of the Thornloe Formation are considered stratigraphically equivalent to the Amabel Formation of southern Ontario (Russell 1984; Chart D, Map 2582, map case). The Thornloe Formation is substantially thicker (*see* Table 20.5; Russell 1984) than the Amabel strata on the Bruce Peninsula (Johnson et al. 1985) or Manitoulin Island (Johnson and Telford 1981a, 1981b), which suggests that carbonate deposition was widespread over the Precambrian shield during this time (Figure 6, Map 2582, map case).

Biohermal and associated platformal carbonates, which characterize Sequence 5 in southern Ontario, also occur throughout the Hudson Platform (Figure 6 and Chart D, Map 2582, map case) and are represented by the Ekwan River Formation and the overlying, and in part, laterally equivalent Attawapiskat Formation (Sanford et al. 1968; Sanford 1987). The Ekwan River Formation, identified in both the Moose and Hudson Bay basins, consists of locally bioclastic or bituminous limestones and dolostones, which are locally biostromal. The Ekwan River Formation is considered to be equivalent to the Fossil Hill Formation and much of the Amabel Formation (Chart D, Map 2582, map case; *see* Figure 20.24; Sanford et al. 1968).

The Attawapiskat Formation consists of biohermal dolostones and limestones with associated bioclastic flank beds and interbiohermal dolostones and limestones (Sanford et al. 1968; Sanford 1987). Parts of this unit have not been dolomitized and show excellent preservation of fauna and sedimentary features. Attawapiskat bioherms are well exposed along the Attawapiskat and Ekwan rivers, on the north side of the Moose River Basin, and along the Severn River, on the south side of the Hudson Bay Basin (Chow and Stearn 1989).

The regional distribution of the Attawapiskat and Ekwan River formations (Figure 6, Map 2582, map case; Sanford 1987) suggests that, as with the Michigan Basin, basin subsidence controlled biohermal development and carbonate facies differentiation around the Hudson and Moose River basins. Thick bank and barrier reef deposits of the Attawapiskat Formation completely encircled the Hudson Bay Basin and fringed the northern side of the Moose River Basin. Significant biohermal build-ups also occurred in central Hudson Bay Basin in response to faulting and uplift of the Central Hudson Bay Arch (*see* Figure 20.30; Sanford 1987; Sanford and Grant 1990).

The close of Sequence 5 deposition is not marked by a widespread, unequivocal unconformity, as are most of the other sequence boundaries. The sequence boundary is, however, marked by a significant change in sedimentation style. In southern Ontario, this is represented in the Guelph to Salina transition and, in northern Ontario, the Attawapiskat to Kenogami River Formation (of Sequence 6) transition. As mentioned in the overview to this sequence, cratonic uplift, probably related to the Salinic Disturbance (a possible early phase of the Acadian Orogeny: Rodgers 1970), restricted the cratonic basins, increasing their salinity and causing deposition of the evaporites of Sequence 6. The sparse fauna in these strata makes the precise timing and interbasin correlation of this event difficult.

## Economic Geology

The Lockport, Amabel and Guelph formations are all major sources of crushed stone in Ontario. As overburden stripping ratios are important in aggregate production, quarries extracting material from these units are mostly located in thin drift areas along the edge of the Niagara Escarpment (Derry, Michener, Booth and Wahl and OGS 1989a, 1989c). The massive-bedded Wiarton/Colpoy Bay Member of the Amabel Formation is quarried for dimension stone on the Bruce Peninsula (Armstrong 1988; Derry, Michener, Booth and Wahl and OGS 1989c). The new Canadian Chancellery in Washington D.C. was recently clad in this material.

The Eramosa Member, on the Bruce Peninsula, is best known as a source of building and landscaping stone (Armstrong and Meadows 1988). Several quarries located on the Bruce Peninsula between Lion's Head and Owen Sound produce a variety of flag, paving and ashlar products, in addition to cut and polished dimension stone (*see* Figure 20.27; Armstrong and Meadows 1988; Derry, Michener, Booth and Wahl and OGS 1989c). The dimension stone products are most decorative if cut and polished parallel to the bedding planes.

Geochemical studies indicate that the Eramosa Member contains total organic carbon contents of up to 3% (Powell et al. 1984; Tworo 1985). Although this is below the limit for "oil shale", it does suggest that this unit may have been a hydrocarbon source rock for conventional oil and gas in other Silurian units (Powell et al. 1984).

About half of the oil production to date in Ontario has come from Silurian rocks. Most of this oil has come from traps in the reefal carbonate rocks of the Guelph Formation (Koepke and Sanford 1966; Bailey and Cochrane 1990) and associated overlying carbonate rocks of the Salina Formation (*see* Sequence 6 for production figures; Carter 1990).

The sandstone of the Thorold Formation, as previously mentioned (*see* Sequence 4), combined with the underlying Grimsby Formation, is a major natural gas source in the Niagara Peninsula area and beneath eastern Lake Erie (*see* Economic Geology section of Sequence 4; Sanford 1969; Brigham 1971; Bailey and Cochrane 1986).

# DEPOSITIONAL SEQUENCE 6: LATE SILURIAN (MIDDLE LUDLOW) TO EARLY DEVONIAN (GEDINNIAN)

## Overview

In Ontario, Sequence 6 is characterized by evaporite deposition with associated carbonate and clastic sediments (mainly orogen-derived clastic sediments in the Western St. Lawrence Platform and craton-derived clastic sediments in the Hudson Platform). The sequence is delimited by a lower boundary, marked by a change from carbonate to evaporite deposition (locally unconformable), and an upper boundary, marked by an extensive erosional unconformity. Both the conditions promoting evaporite deposition and the erosional termination of this sequence, are thought to be products of epeirogenic uplift, brought on by an early phase of the Acadian Orogeny (Sanford, in press).

Widespread regression of marine seaways on the eastern part of the North American craton during the early part of this sequence was apparently caused by epeirogenic uplift, triggered by craton-margin tectonism (Sanford, in press). Rodgers (1970) termed this possible early phase of the Acadian Orogeny, the Salinic Disturbance. This activity produced orogen-derived red beds (Bloomsburg Delta) in northeastern United States, which grade laterally into marine clastic rocks and evaporites in the Appalachian Basin (Rickard 1969), and dolostones, marine shales and thick evaporite deposits (Salina Formation) of the Michigan Basin (Figure 7, Map 2582, map case; Landes 1945a; Sanford 1969). The accelerated subsidence of the Michigan Basin, initiated during Sequence 5 time, continued during this sequence (indicated by isopach contours in Figure 7, Map 2582, map case; Landes 1945a; Sloss 1988). In the Western St. Lawrence Platform, Sequence 6 is capped by normal to semirestricted carbonate deposits (Bertie and Bass Islands formations) of latest Silurian age (Sanford 1969). No sedimentary rocks of Gedinnian age (earliest Devonian) are known in southern Ontario. If deposited, they were subsequently eroded during the regression which terminated this sequence in the Western St. Lawrence Platform.

Limestone fragments of Gedinnian age (Helderberg Formation, not shown on Chart D, Map 2582, map case), representing the closing phase of Sequence 6, are preserved in a Cretaceous(?) diatreme on Sainte-Helene Island at Montreal (Boucot et al. 1986). The older units of this sequence (Ludlow to Pridolian) are assumed to have been deposited in the Ottawa Embayment and Quebec Basin as well, but may have been overlooked in the breccia because of the unfossiliferous character of these units.

The extensive seaway of Sequence 3 to 5 age that connected the Hudson Platform with the St. Lawrence Platform, via the Williston Basin, was severed during Sequence 6 time (Sanford 1987). Thick craton-derived red bed clastic carbonate and evaporite (halite, gypsum-anhydrite) sediments (Kenogami River Formation) were deposited in the Hudson Bay and Moose River basins. The evaporites were largely removed by dissolution. Both of these basins were undergoing accelerated subsidence during this time (Sanford and Grant 1990). The presence of thick craton-derived red bed clastic rocks in both Moose River and Hudson Bay basins indicates widespread uplift of the major arch systems bordering and intersecting the Hudson Platform (Sanford 1987). Perhaps the most intensive, vertical movements occurred (beginning in the Ludlow) along the northwest-trending Central Hudson Bay Arch (Figure 20.30; Roksandic 1987; Thorpe 1988; Sanford and Grant 1990). Parts of this arch, which divided the Hudson Bay Basin into 2 subbasins, remained emergent throughout Ludlow and into early Givetian (Middle Devonian) time.

Details of the stratigraphy of Sequence 6 are described in the following section; a stratigraphic column showing the units comprising this sequence is given in Figure 20.31. The lithologic details of these units are presented in Table 20.6. The present distribution of Sequence 6 units is shown on Maps 2541 to 2544. Their pre-erosion extent is shown on Figure 7, Map 2582 (map case). The distribution of the sequence as a whole is shown on Maps 2575 to 2578[2].

## Geology

### SOUTHERN ONTARIO

Late Silurian orogenic activity, possibly related to the Acadian Orogeny (Salinic Disturbance of Rodgers 1970), was sufficient to cause buckling (i.e., arch uplift and basin subsidence) of the eastern North American craton, but did not appear to introduce much orogen-derived clastic material into southern Ontario (Sanford, in press). This buckling of the craton resulted in the development of restricted circulation in the inland seas and led to widespread deposition of evaporites, associated carbonate sediments and primarily craton-derived clastic sediments.

The initial deposits of Sequence 6 in southern Ontario consist of evaporites, evaporitic carbonates and shales of the Salina Formation (Chart D, Map 2582, map case; *see* Figure 20.31). As previously discussed in Sequence 5, the nature of the lower contact of the Salina Formation with the reefal dolostones of the Guelph Formation is problematic. The Salina Formation is subdivided into 8 lithostratigraphic units, after Landes (1945a) and Sanford (1969), as shown in Figures 20.31 and 20.32 and Table 20.6.

In the Appalachian Basin, the Salina Formation is generally more argillaceous and dolomitic, as compared to the Michigan Basin, where carbonates and evaporites predominate (*see* Figure 20.32). Evaporite deposition in the Appalachian Basin appears to have started later than in the Michigan Basin. Although the thickness and distribution of the individual Salina units are reasonably well known from petroleum exploration activity (Koepke and Sanford 1966; Sanford 1977; Bailey and Cochrane

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[2] *Note: the usage of rock stratigraphic terms Lower, Middle and Upper Silurian is discussed in the Silurian Terminology section in Sequence 4.*

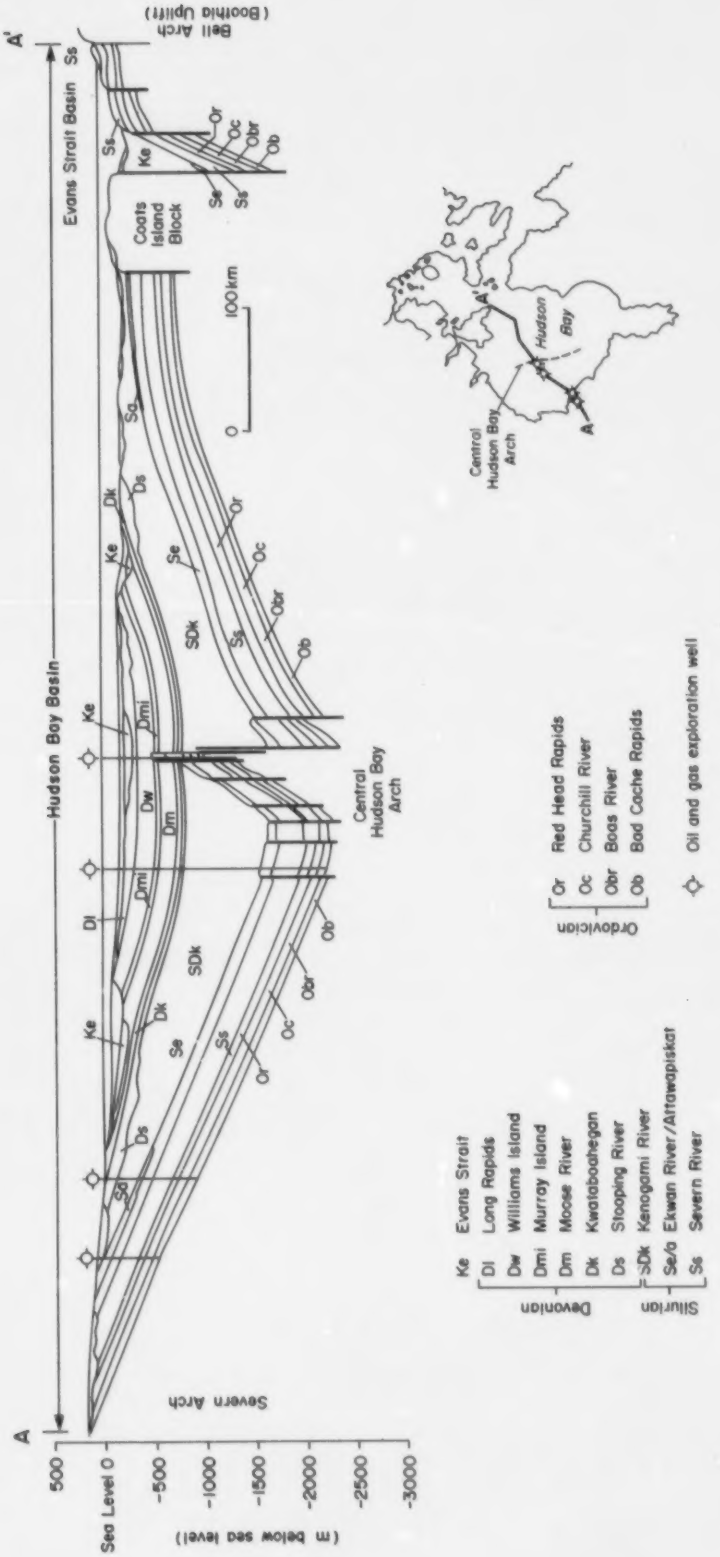
A
A'
Hudson Bay Basin
Evans Strait Basin
Severn Arch
Bell Arch (Boothia Uplift)
Central Hudson Bay Arch
Coats Island Block
500
Sea Level 0
-500
-1000
-1500
-2000
-2500
-3000
(m below sea level)
0
100km
Ke
Dl
Dw
Dmi
Dm
Dk
Ds
SDk
Sa
Se
Ss
Or
Oc
Obr
Ob
Hudson Bay
Central Hudson Bay Arch
Devonian
Ke Evans Strait
Dl Long Rapids
Dw Williams Island
Dmi Murray Island
Dm Moose River
Dk Kwataboahegan
Ds Stooping River
Silurian
SDk Kenogami River
Se/a Ekwan River/Attawapiskat
Ss Severn River
Ordovician
Or Red Head Rapids
Oc Churchill River
Obr Boas River
Ob Bad Cache Rapids
Oil and gas exploration well

**Figure 20.30.** Structure-stratigraphic section across Hudson Bay and Evans Strait basins (*modified from* Sanford and Grant 1990).

**Table 20.6.** Late Silurian (middle Ludlow) to Early Devonian (Gedinnian) units in Ontario—Depositional Sequence 6 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Southern Ontario** | | | | | | |
| Bass Islands Formation | Bruce Peninsula + subsurface of southwestern Ontario | Pridolian | Intertidal to supratidal environments | 22 to 28 m | Dark brown to buff, fine- to microcrystalline dolostone; 2 oolitic dolostone marker beds up to 1 m thick; bituminous streaks; evaporite mineral molds; sparsely fossiliferous (brachiopods, ostracods, stromatolites).<br>Upper contact is unconformable with the Bois Blanc Fm. (Sequence 7).<br>Lower contact with Salina Fm. is conformable. | Sanford 1969; Sparling 1970; Liberty and Bolton 1971; Winder and Sanford 1972 |
| Bertie Formation | Niagara Peninsula | Pridolian | Similar to Bass Islands environment | up to 14 m | Dark brown to buff, finely laminated, argillaceous dolostones with abundant bituminous partings; top surface contains joints or fractures filled with Oriskany Fm. (Sequence 7) sand.<br>Five members identified in Fort Erie area (in ascending order): Oatka, Falkirk, Scajaquada, Williamsville and Akron Mbs.<br>Three lithological subdivisions west of Niagara Peninsula: (lower) thick-bedded, bituminous dolostone; (middle) grey, argillaceous dolostone; (upper) thin-bedded dolostone, locally bituminous.<br>Upper contact is disconformable with the Bois Blanc Fm. (Sequence 7).<br>Lower contact conformable with Salina Fm. | Telford and Tarrant 1975a, 1975b; Telford and Johnson 1984; Sanford 1969; Kobluk et al. 1977; Caley 1941 |
| Salina Formation | Southwestern Ontario | Ludlow to Pridolian | Subtidal through supratidal to terrestrial environmental settings; including lagoonal, sabkha and coastal salina | up to ~330 m | Evaporites, evaporitic carbonates and shales; subdivided into 8 lithostratigraphic units listed below in descending order from youngest to oldest unit.<br>Evaporites (mainly halite) primarily located in units A–1, A–2, B, D, F. Carbonates and evaporites predominant in the Michigan Basin. More argillaceous and dolomitic in the Appalachian Basin.<br>Lower contact with Guelph Fm. (Sequence 5) conformable to disconformable(?). | |
| G unit | | | | 0 to 12 m | Tan dolostone and anhydrite overlain by grey dolomitic shale. | Sanford 1969; Haynes et al. 1988, 1989 |
| F unit | | | | 30 to 90 m | Dolostone and dolomitic shale, with anhydrite, grey and red shaly interbeds and halite. | Sanford 1969; Haynes et al. 1988, 1989 |
| E unit | | | | 15 to 30 m | Grey to brown dolostone and shaly micritic dolostone, grading upward into grey shale; local anhydrite beds. | Sanford 1969; Haynes et al. 1988, 1989 |

**Table 20.6.** Late Silurian (middle Ludlow) to Early Devonian (Gedinnian) units in Ontario—Depositional Sequence 6 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Salina Formation cont'd | | | | | | |
| D unit | | | | 0 to 15 m | Halite, dolostone and shale grades into anhydrite over the Algonquin Arch. | Sanford 1969; Haynes et al. 1988, 1989 |
| C unit | | | | 22 to 30 m | Dolomitic grey shale and anhydrite grading into green shale and shaly dolostone; frosted quartz grains in upper shales. | Sanford 1969; Haynes et al. 1988, 1989 |
| B unit | | | | 0 to 90 m | Dolomitic shales (Niagara Peninsula) grading westward to white, brown-grey, thick, halite beds with thinner interbeds of dolostone, anhydrite and gypsum; gypsum nodules. | Johnson et al. 1985; Sanford 1969; Haynes and Hughes-Pearl 1990a |
| A–2 unit | | | | 12 to 90 m | Divided into 2 informal subunits:<br>A–2 carbonate (upper): dark grey to black limestone and dolostone; microcrystalline.<br>A–2 evaporite (lower): white, brown-grey, thick beds of halite grading laterally and vertically into anhydrite and gypsum intervals. Small-scale dewatering structures; gypsum nodules. | Johnson et al. 1985; Sanford 1969; Haynes and Hughes-Pearl 1990a |
| A–1 unit | | | | 6 to 48 m | Divided into 2 informal subunits:<br>A–1 carbonate (upper): same as A–2 carbonate unit; intervals of dark greyish brown, argillaceous limestones and dolomites.<br>A–1 evaporite (lower): white, brown-grey, halite, anhydrite, gypsum; halite is the most common and occurs in thick beds in the Michigan Basin; anhydrite and/or gypsum horizons (up to 15 m thick).<br>Stromatolites, rip-up clasts; gysum nodules; dewatering and/or deformation structures. | Johnson et al. 1985; Sanford 1969: Haynes et al. 1988, 1989; Haynes and Hughes-Pearl 1990a |
| **Northern Ontario** | | | | | | |
| Kenogami River Formation | Hudson Bay and Moose River basins | Ludlow to Gedinnian | Upper mb.: intertidal with periods of subaerial exposure.<br>Middle mb.: red beds deposited in supratidal environment with sandstone lenses representing fluvial deposits.<br>Lower mb.: subtidal and intertidal environments. | Upper mb.: 8 to 53 m<br>Middle mb.: 145 to 168 m<br>Lower mb.: 23 to 53 m | Subdivided into 3 members:<br>Upper mb.: tan and cream, microcrystalline and oolitic dolostone with dolostone breccia.<br>Middle mb.: red (with minor green) gypsiferous mudstone, sandstone, and minor argillaceous dolostone and "honeycombed" limestone; local halite occurrences; gradationally overlies lower mb.<br>Lower mb.: brown, fine-crystalline dolostone; thin to thick bedded; interbedded with gypsum and anhydrite; local halite.<br>Evaporites (mainly halite) are more abundant in the Hudson Bay Basin section.<br>Upper contact with overlying Stooping River Fm. (Sequence 7) is unconformable.<br>Lower contact with the Attawapiskat or Ekwan River Fms. (Sequence 5) is conformable(?). | Sanford 1987; Sanford et al. 1968; Sanford and Norris 1973, 1975 |

1990), sedimentological and paleontological details of the units remain relatively poorly understood.

Near the southern end of Lake Huron, evaporites in the Salina Formation have a composite thickness in excess of 200 m, although a thickness of 30 to 60 m is more typical for southwestern Ontario (Sanford 1969). The evaporites are mainly located in the A–1, A–2, B, D and F units and consist of halite, anhydrite and gypsum (*see* Figure 20.32). Halite is the most common evaporite mineral, occurring in very thick beds, particularly in the A and B units in the Michigan Basin (Figure 20.33; Sanford 1977; Johnson et al. 1985).

Halite beds in the Salina Formation are now largely restricted to the Michigan and Appalachian basins proper, having been subjected to post-deposition dissolution elsewhere (*see* Figure 20.32; Figure 7, Map 2582, map case; Brigham 1971); collapse breccias in the overlying units suggests that at least the B Salt was deposited over the Algonquin Arch (Sanford 1965, 1969, 1977). Salt dissolution began at the margins of the Appalachian and Michigan basins and progressed toward the basin centres. Anomalous thicknesses of the overlying Bertie and Bass Islands formations indicate that the dissolution of salt from the Salina Formation began soon after deposition (Sanford 1969). Recent subsurface studies (e.g., Sanford et al. 1985) have suggested that collapse zones resulting from salt dissolution follow a regional joint and/or fault pattern (*see* Summary and Discussion).

A number of anhydrite-gypsum horizons up to 2 m thick are present within the Salina Formation (Figure 20.34); some (e.g., the A–1 and A–2 anhydrites) are used as marker horizons for oil and gas exploration (Beards 1967). In the subsurface, this evaporite is present mainly as anhydrite; however, in subcrop areas, groundwater infiltration has converted the anhydrite to gypsum.

Recent detailed work on gypsum deposits of the Salina Formation, in the vicinity of the Algonquin Arch, has shown

| | **Northern Ontario** | **Southern Ontario** | | | |
|---|---|---|---|---|---|
| **Series** | **Hudson Bay and Moose River basins** | **Michigan Basin** | | **Appalachian Basin (Niagara Peninsula)** | |
| Lower Devonian | Kenogami River Fm. | | | | |
| Upper Silurian | | Bass Islands Fm. | | Bertie Fm. | |
| | | Salina Fm. | G | Salina Fm. | G |
| | | | F | | F |
| | | | E | | E |
| | | | D | | D |
| | | | C | | C |
| | | | B | | B |
| | | | A–2 | | A |
| | | | A–1 | | |

**Figure 20.31.** Stratigraphic units of Sequence 6.

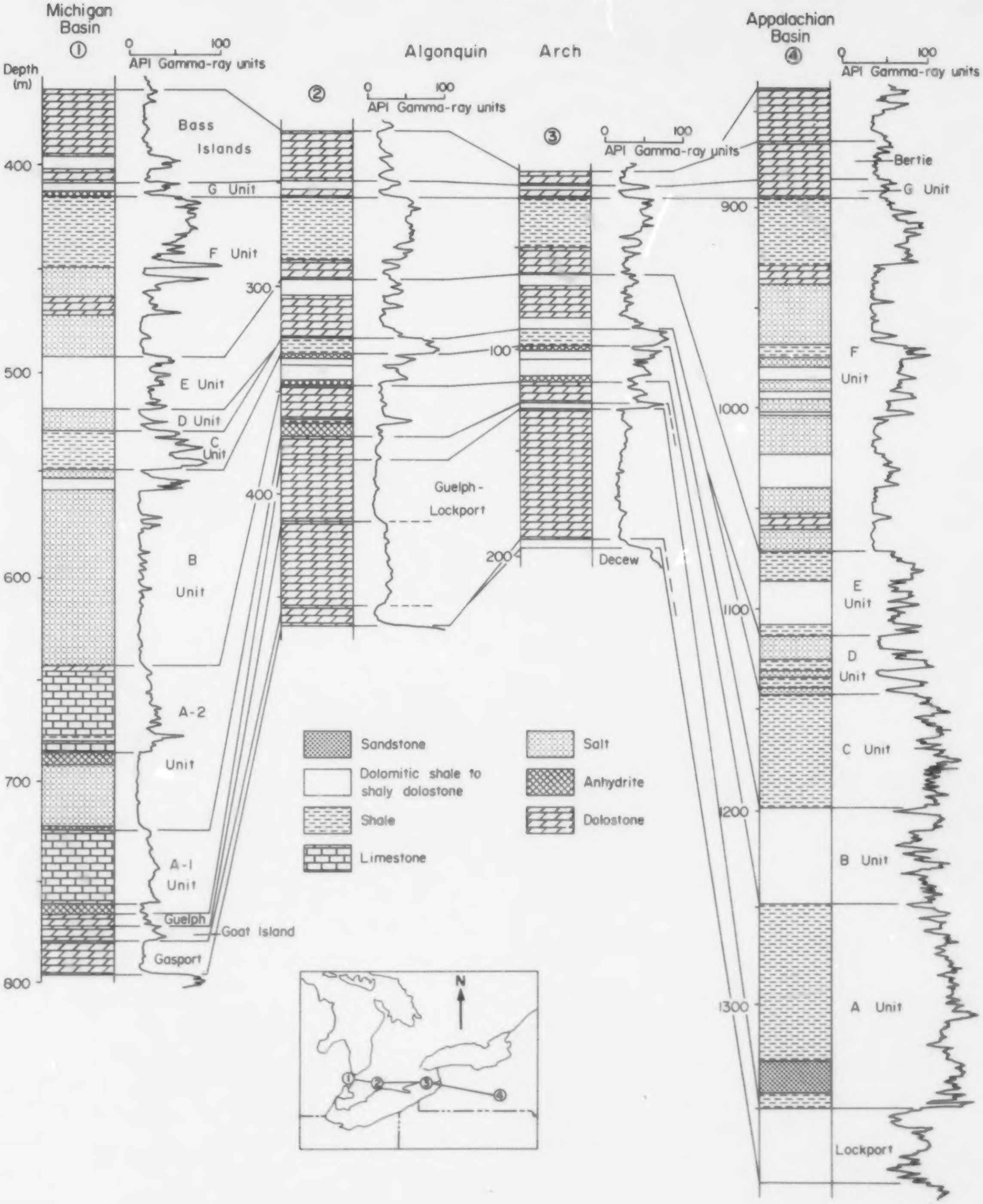
Michigan Basin
①
0
100
API Gamma-ray units
Depth (m)
400
500
600
700
800
Bass Islands
G Unit
F Unit
E Unit
D Unit
C Unit
B Unit
A-2 Unit
A-1 Unit
Guelph
Goat Island
Gasport
Algonquin Arch
②
0
100
API Gamma-ray units
300
400
③
0
100
API Gamma-ray units
100
200
Guelph-Lockport
Decew
Appalachian Basin
④
0
100
API Gamma-ray units
Bertie
G Unit
900
1000
1100
1200
1300
F Unit
E Unit
D Unit
C Unit
B Unit
A Unit
Lockport
Sandstone
Dolomitic shale to shaly dolostone
Shale
Limestone
Salt
Anhydrite
Dolostone
N
①
②
③
④

**Figure 20.32.** Stratigraphic correlation of Upper Silurian (Sequence 6) strata across southwestern Ontario, from the Michigan Basin to the Appalachian Basin (*modified from* Carter 1987).

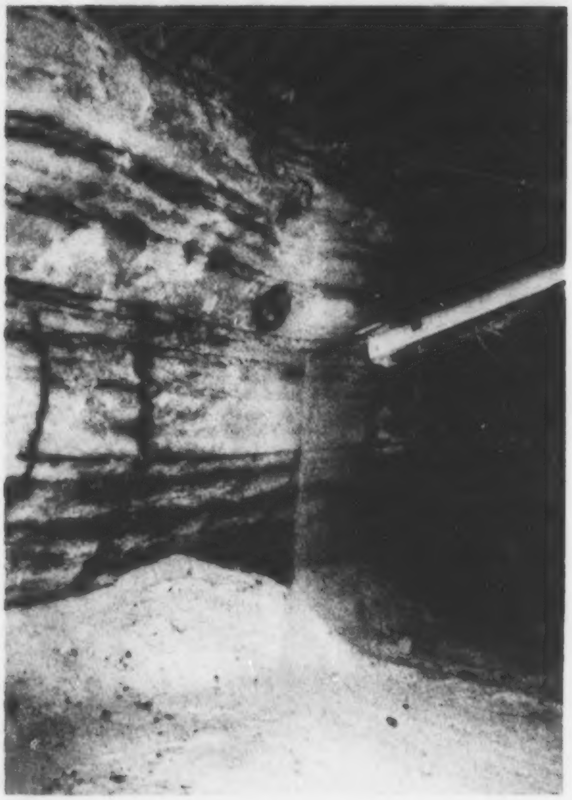

**Figure 20.33.** The Salina Formation "A–2" salt bed in the Sifto Canada Inc. salt mine at Goderich (photograph *courtesy of* B. Feenstra).

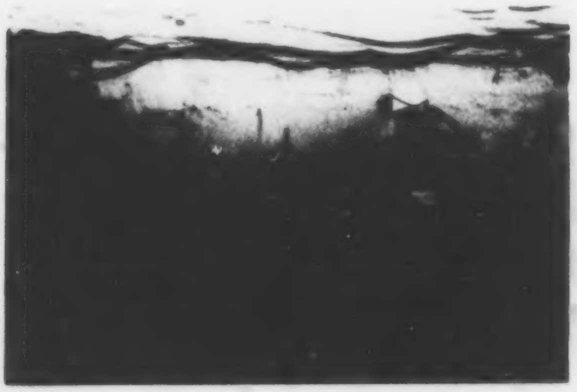

**Figure 20.34.** "Main" gypsum bed (1.7 m thick) in the Westroc Mine at Drumbo, northeast of Woodstock. This is the basal gypsum-anhydrite bed in the A–2 unit of the Salina Formation (photograph *courtesy of* R. Bezys).

them to be vertically cyclical and laterally complex (Haynes et al. 1988, 1989; Haynes and Hughes-Pearl 1989, 1990a, 1990b). These workers proposed depositional environments ranging from subtidal through supratidal to terrestrial, in settings ranging from lagoonal through sabkha to coastal salina, and have suggested modern analogs of Egypt and the Sinai Peninsula for the Salina Formation in the Algonquin Arch area. A sabkha environment has also been proposed for the Salina Formation in New York State (Treesh and Friedman 1974). Depositional environments ranging from deep, stratified, silled or restricted basin through to subaerial sabkha have been proposed for the Salina Formation in the Michigan Basin (e.g., Nurmi and Friedman 1977).

The Salina Formation is conformably overlain by dolostones of the Bertie and Bass Islands formations (*see* Table 20.6 and Figure 20.31; Chart D, Map 2582, map case), the youngest Silurian strata in southern Ontario. The Bertie Formation ("Bertie–Akron" of Caley 1941) is an Appalachian Basin unit, restricted in Ontario to the Niagara Peninsula (Liberty and Bolton 1971). The laterally equivalent Bass Islands Formation, on the other hand, is a Michigan Basin unit, with its type section located on the Bass Islands in Lake Erie (Sparling 1970; Liberty and Bolton 1971). The lateral transition from the Bertie Formation to the Bass Islands Formation is gradational, occurring north and west of Wilsonville, which is situated approximately 10 km south of Brantford (Telford 1979; Telford and Hamblin 1980).

The Bertie Formation outcrops sporadically in the Niagara Peninsula, particularly near the Niagara River. The resistant dolostones of this unit form the mostly buried Onondaga escarpment, which was formed by differential erosion of the softer, underlying Salina Formation (Map 2544, map case; Feenstra 1974; Telford and Tarrant 1975a, 1975b; Chapman and Putnam 1984). The Onondaga escarpment parallels the Niagara Escarpment from Fort Erie to Hagersville, west of which its presence is obscured by glacial cover.

In the Fort Erie area of the Niagara Peninsula, 5 members have been identified in the Bertie Formation (*see* Table 20.6; Telford and Johnson 1984). West of the Niagara Peninsula, these subdivisions are less distinct, and the Bertie Formation is subdivided into as many as 3 informal lithologic units (*see* Table 20.6; Telford and Tarrant 1975a, 1975b). Generally, the Bertie Formation is a relatively dark coloured dolostone, characterized by numerous bituminous partings.

The Bass Islands Formation is poorly exposed in its subcrop area, which extends across Ontario from just south of Brantford, in a north-northwest direction, to Port Elgin on Lake Huron (Map 2544, map case; Sanford 1969). The unit is generally a dark dolostone and is unsubdivided in Ontario because of its homogeneous nature and poor exposure. Liberty and Bolton (1971) reported 2 oolitic marker beds in the Bass Islands Formation, in the subsurface south of the Bruce Peninsula.

The top of the Bertie and Bass Islands formations (and Sequence 6 in southern Ontario) is an erosional surface marking a major regional unconformity. Sand and pebbles, of Devonian age, are commonly present in cracks and joints extending downward for several metres from the top of the Bertie and Bass Islands strata (*see* Sequence 7; Kobluk et al. 1977; Telford and Johnson 1984; Derry, Michener, Booth and Wahl and OGS 1989c).

The maximum combined thickness of the Salina and Bass Islands formations in southern Ontario ranges from

518 m, near southern Lake Huron, to only 105 m over the Algonquin Arch (Winder and Sanford 1972). In a drill hole near Sarnia, the Salina Formation is 330 m thick and the Bass Islands Formation is 24 m thick (Johnson et al. 1985).

The relatively unfossiliferous nature of the Upper Silurian formations in Ontario renders age determinations and regional correlation of these units far less precise than for either Ordovician or Devonian strata. Correlation of the Salina and Bertie and Bass Islands formations is based mostly on the stratigraphic position of these units (*see* Winder et al. 1975; Telford 1978; and discussion in Sequence 4).

### NORTHERN ONTARIO

In the Hudson Platform, sedimentary deposits of Sequence 6 are represented by the red bed siliciclastic rocks, carbonates and evaporites (mainly halite) of the Kenogami River Formation (Chart D, Map 2582, map case; *see* Figure 20.31). This unit reaches a maximum thickness of about 800 m in the Hudson Bay Basin, whereas in the Moose River Basin, it attains a thickness of only 250 m (Sanford 1987). The Kenogami River Formation is better known in the Moose River Basin where there is a greater amount of drill hole data.

In both the Hudson Bay and Moose River basins, the Kenogami River Formation can be separated into 3 informal members (*see* Table 20.6; Sanford et al. 1968; Sanford and Norris 1975). The dolostones and minor evaporites of the lower member were deposited within subtidal and intertidal environments (Sanford and Norris 1973). The clastic and carbonate rocks of the middle member reflect a change to craton-derived red-bed sedimentation in a supratidal setting, with local fluvial deposition represented by sandstone lenses (Sanford and Norris 1973). The upper member, consisting of oolitic and brecciated dolostone, represents a return to intertidal conditions with periods of subaerial exposure (Sanford and Norris 1975). The presence of breccia in the upper member is likely a product of evaporite dissolution and collapse.

The Kenogami River Formation spans the Silurian–Devonian boundary. The system boundary occurs within the middle member, close to the middle-upper member contact (McGregor and Camfield 1976). The reader should note that this makes the system boundary shown on the bedrock geology maps (Maps 2541 to 2543) somewhat arbitrary. No unconformity comparable to the regional Gedinnian unconformity marking the end of Sequence 6 in southern Ontario (Chart D, Map 2582, map case) is known in the Hudson Platform.

The top of the Kenogami River Formation marks the beginning of a major hiatus equivalent to much of the Siegenian stage (McGregor and Camfield 1976) and the end of Sequence 6 deposition in the Hudson Platform.

## Economic Geology

The Salina Formation is a major source of salt, brine, gypsum, oil and gas in Ontario. Total production of salt from this unit in Ontario amounted to 5.7 million tonnes in 1987 (Ministry of Northern Development and Mines 1988) and 6.7 million tonnes in 1988 (Prud'homme 1989). Rock salt is mined by underground methods at Windsor by The Canadian Salt Company Ltd. and at Goderich by Sifto Canada Inc. (*see* Figure 20.33). Combined production amounted to 4.8 million tonnes during 1988 (Prud'homme 1989). Most of this rock salt is used in de-icing. Salt is also mined in Ontario by solution mining methods. Solution mining has been applied in southern Ontario since 1880 (Prud'homme 1989; Ministry of Northern Development and Mines 1988); production amounted to about 1.8 million tonnes in 1988 (Prud'homme 1989). This mining method involves the injection of fresh water into salt beds at depth and the subsequent collection of brines through the same or adjacent wells. Solution mining of the Salina Formation is conducted at Amherstburg by General Chemical Canada Inc. and at Sarnia by Dow Chemical Canada Inc. Brines produced there are mostly used in industrial chemical applications. Smaller-scale solution mining operations at Windsor and Goderich produce fine salt for food grade salt, the chemical industry and agricultural uses (Trevail et al. 1986). Large subsurface caverns produced by solution mining are believed to be responsible for local surface subsidence in the Windsor–Detroit area (e.g., Nieto et al. 1983; discussed in Sequence 7).

Gypsum in the Salina Formation is mined from subsurface operations at Hagersville (Canada Gypsum Ltd.), Caledonia (Domtar Construction Materials Ltd.) and Drumbo (Westroc Industries Ltd.; *see* Figure 20.34) (Haynes et al. 1988, 1989; Haynes and Hughes-Pearl 1989, 1990a). In 1987, these deposits produced about 1.5 million tonnes of gypsum (Haynes et al. 1989). Most of this gypsum is used to manufacture wallboard. The proximity of these deposits to the lucrative southern Ontario market has enhanced their economic viability.

Hydrocarbons from the Salina Formation occur mainly in structural (i.e., dome) traps caused by the draping of Salina strata (usually restricted to the A–1 and A–2 carbonate units) over Guelph Formation reefs (Bailey and Cochrane 1990). As the production from these traps is often combined with that from the underlying reefs, the value of production from Salina traps is unknown. In 1989, oil production from the combined Guelph and Salina formations amounted to 27 262 $m^3$, accounting for 11% of the total oil production in Ontario. Gas production from these same 2 units totalled 168 516 thousand $m^3$, accounting for 34.5% of the total Ontario production. Cumulative oil and gas production from these units through to the end of 1989 was 1 883 000 $m^3$ and 17 474 000 thousand $m^3$, respectively, accounting for 19% and 61% of the total Ontario production, respectively (Carter 1990).

The Bertie Formation is quarried at a number of locations between Fort Erie and Hagersville, for a variety of crushed stone products (Derry, Michener, Booth and Wahl and OGS 1989c). The Bass Islands Formation is too poorly exposed to be of economic interest to the limestone industry.

# DEPOSITIONAL SEQUENCE 7: LATE EARLY DEVONIAN (SIEGENIAN) TO EARLY MIDDLE DEVONIAN (EIFELIAN)

## Overview

Sequence 7 records 2 transgressive episodes subsequent to the major Early Devonian unconformity which forms its lower boundary. The first, brief transgression (during the Siegenian) is represented by thin, erosional remnants of originally widespread craton-derived clastic sediments. The second and principal transgression (Emsian and Eifelian age) deposited minor craton-derived clastic sediments, basin-centred evaporites and extensive carbonate sediments from the St. Lawrence to Hudson platforms. Sequence 7 was terminated by a major regional unconformity caused by uplift and regression related to the onset of the Acadian Orogeny (*see* Sequence 8).

Epeirogeny, related to early Acadian tectonism, provided a renewed source of craton-derived detritus during early Sequence 7 time. These sediments were deposited during the Siegenian age transgression, which extended over a wide part of the eastern North American craton (Sanford, in press). These Siegenian age beds, which, in Ontario, consist of sandstones (Oriskany Formation), were subjected to subaerial erosion immediately following deposition. Although widespread in the Appalachian Basin, only minor erosional remnants occur in southwestern Ontario and Michigan (Landes et al. 1945; Oliver et al. 1968; Winder and Sanford 1972). Fragments of this deposition are also preserved in the Sainte-Helene Island breccia at Montreal (Boucot et al. 1986).

The second and principal transgression of Sequence 7 occurred during the Emsian to Eifelian and was substantially more widespread, forming a direct seaway between southern Ontario and the Hudson Platform (Sanford, in press). In the Western St. Lawrence Platform, this second trangression consists mainly of carbonates and evaporites, with minor basal orthoquartzitic sandstones, lying unconformably upon Upper Silurian strata of Sequence 6 (Chart D, Map 2582, map case; Cohee and Underwood 1945; Landes 1945b; Oliver 1966; Sanford 1968; Uyeno et al. 1982). This succession is complex and not well understood due to its relatively poor exposure (beneath thick Quaternary deposits) and rapid lateral facies changes. The latter was caused by irregular depositional surfaces formed by differential subsidence, related to continued dissolution of underlying Silurian evaporites (Brigham 1971; Sanford et al. 1985).

In the Central St. Lawrence Platform, limestone fragments of Emsian to Eifelian age have been identified in the Sainte-Helene Island breccia at Montreal (Boucot et al. 1986), but elsewhere on the platform, Sequence 7 strata are absent (Figure 8, Map 2582, map case).

In the Hudson Platform, Emsian to Eifelian deposition is represented by strata which are lithologically and faunally similar to their counterpart units in southern Ontario and Michigan (Chart D, Map 2582, map case; Sanford and Norris 1975; Telford 1988). In the Hudson Bay Basin, Sequence 7 strata rest with an unconformable contact on Sequence 6 rocks (Kenogami River Formation). In the Moose River Basin, Sequence 7 strata similarly rest on Sequence 6 rocks, and in the southern part of the basin, they overlap these strata to rest on the uplifted edges of Lower Silurian and Upper Ordovician strata or on Precambrian crystalline rocks of the Fraserdale Arch (Sanford and Norris 1975; Sanford 1987). Evaporites deposited during this sequence (mainly in the Moose River Formation) in both Hudson Bay and Moose River basins (Sanford and Norris 1975; Sanford 1987) are considered approximately equivalent to lithologically similar units (Lucas Formation) of the Michigan Basin (Landes 1951; Sanford 1968).

The upper boundary of Sequence 7 is marked, in both the St. Lawrence and Hudson platforms, by a regional-scale unconformity, sharply overlain by orogen-derived clastic rocks, which signal the onset of the main Acadian Orogeny (*see* Sequence 8).

The stratigraphy of Sequence 7 is described in detail in the following section and illustrated in Figure 20.35. Lithologic details of the units included in this sequence are provided in Table 20.7. The present areal distribution of these units is shown on Maps 2541 to 2544 (map case). Their past and present distributions are illustrated in Figure 8, Map 2582 (map case). The distribution of the sequence is shown on Maps 2575 to 2578 (map case).

## Geology

### SOUTHERN ONTARIO

The Early Devonian in southern Ontario was characterized by an extended period of erosion. The only strata known to have been deposited during this time were the Siegenian (approximate) age, shallow marine sandstones of the Oriskany Formation (*see* Figure 20.35). The duration of the erosional episodes preceding and following deposition of this unit are not known precisely (Chart D, Map 2582, map case).

The lithostratigraphic boundary between the Oriskany Formation and underlying Silurian Bertie Formation is not well exposed in Ontario; however, where observable (e.g., Cayuga Quarry), the contact is generally sharp and disconformable, showing pronounced small-scale karst features (Kobluk et al. 1977; Derry, Michener, Booth and Wahl and OGS 1989c).

The Oriskany Formation is a coarse-grained, sparsely fossiliferous, calcareous quartz sandstone with a thin basal conglomerate (Stauffer 1915; Derry, Michener, Booth and Wahl and OGS 1989c). This relatively thin formation is confined to an erosional depression and restricted in both outcrop and subcrop to a small area (approximately 6 km²) 10 km east of Hagersville on the Niagara Peninsula (Sanford 1968; Uyeno et al. 1982). It pinches out laterally between the underlying Bertie Formation and the overlying Bois Blanc Formation (Stauffer 1915; Uyeno et al. 1982).

Where the Oriskany Formation is not present, the Bertie Formation is overlain directly by the Bois Blanc Formation. In these areas, some joints in the Bertie Formation are filled with Oriskany sand, suggesting that the Oriskany Formation was once more extensive, and that it had been subjected to erosion during the Emsian transgression prior to deposition of the overlying carbonate sediments (Kobluk et al. 1977; Sanford 1968; Chart D, Map 2582, map case).

In southern Ontario, the bulk of Sequence 7 consists of relatively pure carbonates with minor evaporites, sandstones and clastic-rich carbonates (*see* Table 20.7; Chart D, Map 2582, map case). Most of the sandstones are associated with early phases of transgressive episodes. The carbonate rocks are predominantly limestone; however, most become more dolomitic toward the central part of the Michigan Basin. Regionally, this mainly carbonate sequence thins across southern Ontario, from approximately 150 m in the Windsor area to approximately 100 m in the Niagara Peninsula area (Uyeno et al. 1982). In addition, most of the units in the sequence thin over the Algonquin Arch. The arch influenced deposition of the lower carbonate units of this sequence, resulting in the interfingering of units typical of the Appalachian Basin (e.g., Onondaga Formation) with those units typical of the Michigan Basin (e.g., Lucas Formation) (Uyeno et al. 1982; Johnson et al. 1989).

The Bois Blanc Formation, which forms the base of Sequence 7 where the Oriskany Formation is absent, is

| | Northern Ontario | Southern Ontario | | | | |
|---|---|---|---|---|---|---|
| **Series** | **Hudson Bay and Moose River basins** | **Michigan Basin** | | | **Appalachian Basin** | |
| Middle Devonian | Murray Island Fm. | Dundee Fm. | | | Dundee Fm. | Marcellus Fm. |
| | Moose River Fm. | Detroit River Gp. | Lucas Fm. | Anderdon Mb. | Onondaga Fm. | Moorehouse Mb. |
| | Kwataboahegan Fm. | | Amherstburg Fm. | | | Clarence Mb. |
| | | | Sylvania Fm. | | | Edgecliffe Mb. |
| Lower Devonian | Stooping River Fm.<br>Sextant Fm. | Bois Blanc Fm. | | | Bois Blanc Fm. | |
| | | | | | | Springvale Mb. |
| | | | | | Oriskany Fm. | |

**Figure 20.35.** Stratigraphic units of Sequence 7.

**Table 20.7.** Early Devonian (Siegenian) to Middle Devonian (Eifelian) units in Ontario—Depositional Sequence 7 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Southern Ontario** | | | | | | |
| Marcellus Formation | Southwestern Ontario (Lake Erie to St. Thomas area) | Eifelian | Marine environment with stratified water column | up to 12 m | Black, organic-rich shale with interbeds of grey shale and very fine- to medium-grained, impure carbonates. Grey shale beds contain brachiopods; carbonate beds contain echinoderms and brachiopod debris; black shale is sparsely fossiliferous with locally abundant spore cases.<br>Two distinct organic-rich beds with organic contents of 7.38% and 4.14%, respectively.<br>Upper contact is sharp and disconformable with Bell Fm. of Hamilton Gp. (Sequence 8).<br>Lower contact is conformable with Dundee Fm. | Johnson et al. 1989;<br>Brown 1985;<br>Telford 1984 |
| Dundee Formation | Southwestern Ontario (Lake Erie to Lake Huron and Lake St. Clair) | Eifelian | Environments range from lagoonal to open shelf and deep water | 35 to 45 m (avg.) | Brown, medium- to thick-bedded, fossiliferous limestone with bituminous partings and chert nodules; organic-rich beds in uppermost metre; brachiopods, molluscs, bryozoans, trilobites and corals. Locally, bitumen infilling vugs and corals. Hardgrounds or firmgrounds locally developed. More massive and fossiliferous, lighter brown and bioclastic on Pelee Island. Thinner bedded with horizons of bituminous limestone on the Niagara Peninsula.<br>Upper contact (with Hamilton Gp.) and lower contact (with Lucas or Amherstburg Fms.) are sharp and erosional. | Uyeno et al. 1982;<br>Sanford 1968;<br>Johnson et al. 1985;<br>Derry, Michener, Booth and Wahl and OGS 1989c;<br>Birchard 1990;<br>Hamilton and Coniglio 1990 |
| Lucas Formation (Detroit River Group) | Southwestern Ontario (excluding Niagara Peninsula area) | Eifelian | Very shallow marine to evaporitic | 40 to 75 m | Light grey to brown, thin- to medium-bedded, limestone and dolostone; commonly bituminous and cherty; locally thin anhydrite and/or gypsum partings and blebs. Needle-like moldic porosity; locally brecciated.<br>In central Michigan Basin: light brown to grey-brown limestone and dolostone alternating with anhydritic dolostone, anhydrite and halite.<br>Anderdon Mb. (upper mb.): light to dark grey-brown, fine-grained (micritic), high-calcium limestone alternating with coarse-grained bioclastic limestone; includes cream to tan, sandy limestone unit ("Columbus Sands").<br>Lower contact is conformable with Amherstburg Fm. | Sanford 1968;<br>Uyeno et al. 1982;<br>Johnson et al. 1985;<br>Telford and Hamblin 1980;<br>Birchard 1990 |

**Table 20.7.** Early Devonian (Siegenian) to Middle Devonian (Eifelian) units in Ontario—Depositional Sequence 7 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Amherstburg Formation (Detroit River Group) | Southwestern Ontario (excluding Niagara Peninsula area) | Emsian to Eifelian | Shallow marine | 40 to 60 m<br>Formosa Reef Limestone: 15 m | In Ingersoll area, tan to grey-brown, fine- to coarse-grained, bioclastic cherty and bituminous limestone; massive-bedded, laminated, bituminous dolostone to southwest, with oolitic, bioclastic and coralgal zones.<br>Formosa Reef Limestone (Bruce and Huron counties): reef complex; grey, fine-grained, pure limestone, dominantly containing stromatoporoids, also includes corals, brachiopods, molluscs, trilobites.<br>Lower contact is conformable with Bois Blanc Fm. and with Sylvania Fm. in southwest. | Best 1953;<br>Fagerstrom 1961;<br>Sanford 1968;<br>Uyeno et al. 1982;<br>Klein 1980;<br>Derry, Michener, Booth and Wahl and OGS 1989a |
| Sylvania Formation (Detroit River Group) | Southwestern Ontario (Windsor to Sarnia area) | Emsian to Eifelian | Strand line and eolian deposits | up to 25 m | White, fine- to medium-grained orthoquartzitic sandstone; also called "Sylvania Sandstone" and "Sylvania Member of Amherstburg Fm." | Sanford 1968;<br>Uyeno et al. 1982;<br>Nieto et al. 1983 |
| Onondaga Formation | Niagara Peninsula | Emsian to Eifelian | Shallow marine | Moorehouse Mb.: 4.5 m<br>Clarence Mb.: ~5 to 8 m<br>Edgecliffe Mb.: 21 m | Divided into 3 members (in descending order):<br>Moorehouse Mb.: dark grey-brown, medium-bedded, biostromal limestone; chert as nodules and thin beds; very fossiliferous.<br>Clarence Mb.: dark grey to brown, thick-bedded, very cherty limestone (chert >50%); sparsely fossiliferous.<br>Edgecliffe Mb.: dark to light grey, medium-bedded, biostromal limestone; grey to black chert nodules; local chert-free, richly fossiliferous argillaceous units that weather into shaly beds; local biohermal mounds with rugose corals. | Telford and Tarrant 1975a, 1975b;<br>Uyeno et al. 1982;<br>Derry, Michener, Booth and Wahl and OGS 1989c |
| Bois Blanc Formation | Southwestern Ontario (Niagara Peninsula to Lake Huron) | Emsian | Shallow marine | 3 to 50 m<br>Springvale Mb.: 3 m avg. (50 m max.) | Green-grey to grey-brown, thin- to medium-bedded, fine- to medium-grained, cherty limestone; grades westward into dolostone; local shaly and glauconitic partings; brachiopods and corals common; locally very fossiliferous.<br>Springvale Mb. (basal unit in Niagara Peninsula area): white, glauconitic, quartz sandstone or sandy carbonate; medium to coarse grained and friable.<br>Lower contact is unconformable with Bertie and Bass Islands Fms. (Sequence 6) or Oriskany Fm. and commonly marked by several metres of relief. | Stauffer 1915;<br>Kobluk et al. 1977;<br>Telford and Tarrant 1975a;<br>Uyeno et al. 1982;<br>Sanford 1968;<br>Derry, Michener, Booth and Wahl and OGS 1989c |

**Table 20.7.** Early Devonian (Siegenian) to Middle Devonian (Eifelian) units in Ontario—Depositional Sequence 7 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| Oriskany Formation | Niagara Peninsula area, near Cayuga and Hagersville | Siegenian | Shallow marine | 6 m (max.) | Yellow to grey, coarse-grained, calcareous quartz sandstone; sparsely fossiliferous; thin (up to 30 cm) basal interval of conglomerate consisting of glauconitic shale, dolostone, mudstone and sandstone clasts. Fossiliferous intervals, mostly brachiopods.<br>Lower contact is sharp and unconformable with Bertie Fm. (Sequence 6) and shows small-scale karst features. | Kobluk et al. 1977; Stauffer 1915; Derry, Michener, Booth and Wahl and OGS 1989c; Sanford 1968; Uyeno et al. 1982 |
| **Northern Ontario** | | | | | | |
| Murray Island Formation | Moose River and Hudson Bay basins | Eifelian | Shallow marine | 6 to 20 m | Light brown, fossiliferous limestone. Fossils include brachipods, crinoids. Halite locally present at base of unit with minor claystone, mudstone and dolostone. Brecciation at base.<br>Lower contact is brecciated and disconformable with Moose River Fm. | Sanford and Norris 1975; Thorpe 1988; Norris 1986; Telford 1988 |
| Moose River Formation | Moose River and Hudson Bay basins | Eifelian | Restricted, very shallow marine to sabkha | 28 to 90 m | Brecciated sparesely fossiliferous limestone; gypsum breccia (clasts and matrix); massive white gypsum; basal sucrosic, vuggy dolostone; local anhydrite; halite (Hudson Bay Basin). Brecciation is extensive; green mudstone matrix is common.<br>Lower contact is conformable with Kwataboahegan Fm. | Bezys 1990; Telford 1988; Thorpe 1988; Norris 1986; Sanford and Grant 1990; Sanford and Norris 1975; Sanford 1987 |
| Kwataboahegan Formation | Moose River and Hudson Bay basins | Emsian to Eifelian | Shelf-edge, reefal and interreefal deposits | 24 to 77 m | Tan to brown, massive- to thick-bedded, very fossiliferous, biostromal and biohermal limestones. Local stromatoporoidal bindstones and crinoidal packstones; local thinner bedded bituminous limestone and dolostone. Fossils include stromatoporoids, crinoids, corals, brachiopods and trilobites.<br>Lower contact is conformable with Stooping River Fm. | Sanford et al. 1968; Thorpe 1988; Telford 1988; Sanford and Norris 1975 |
| Stooping River Formation | Moose River and Hudson Bay basins | Emsian | Intertidal to shallow subtidal environment | <50 m (avg.) (143 m max.) | Grey to tan, cherty limestone with intervals of dolomitic limestone and dolostone; richly fossiliferous. Fossils include brachiopods, corals, echinoderms, gastropods, trilobites and tentaculitids.<br>Lower part grades laterally into Sextant Fm.<br>Lower contact is unconformable with Kenogami River Fm. (Sequence 6) and older units. | Sanford et al. 1968; Telford 1988; Sanford and Norris 1975; Thorpe 1988 |
| Sextant Formation | South margin of Moose River Basin | Emsian | Probably nonmarine origin; possibly marginal marine environment | 90 m (max.) | Red, arkosic sandstone with lesser quartz-feldspar, varicoloured conglomerate, red-grey siltstone, clay and shales; well-preserved plant and spore fossils.<br>Lower contact is unconformable over Precambrian basement. | McGregor and Camfield 1976; Stoakes 1978; Telford 1988; Sanford and Norris 1975; Sanford et al. 1968; Heuber 1983 |

characterized by cherty limestones with a basal sandstone unit (Springvale Member) in the Appalachian Basin. The cherty limestones grade westward into dolostones in the Michigan Basin. Although present on the Algonquin Arch and in the Appalachian Basin, the unit is considered part of the Michigan Basin Devonian succession, thinning eastward from a maximum thickness in Huron County of about 50 m to a minimum thickness of 3 to 15 m on the Niagara Peninsula (Sanford 1968; Uyeno et al. 1982). In subcrop, the Bois Blanc Formation extends in a southeasterly direction from Lake Huron to near Hagersville, and eastward to Fort Erie (Map 2544, map case; Sanford 1968; Uyeno et al. 1982). On the Niagara Peninsula, this unit is relatively well exposed, commonly forming the cap rock of the Onondaga escarpment (Feenstra 1974; Telford and Tarrant 1975a; Chapman and Putnam 1984).

In the Niagara Peninsula area, the basal member of the Bois Blanc Formation is a glauconitic sandstone or sandy carbonate known as the Springvale Member (Telford and Tarrant 1975a). This unit may consist of reworked Oriskany material (Stauffer 1915; Uyeno et al. 1982). Although quite thin (*see* Table 20.7), greater thicknesses (up to 50 m), reflecting infilling of topographic lows in the eroded uppermost Silurian surface have been noted (Sanford 1968). Such local thickening, also observed in the Upper Silurian Bass Islands Formation and in younger Devonian rocks (e.g., Detroit River Group and the Dundee Formation), may be related to the collapse of underlying strata as a consequence of fault-controlled salt dissolution in the Upper Silurian Salina Formation (Sanford 1968; Brigham 1971; Sanford et al. 1985).

The Bois Blanc Formation's lower disconformable contact with the Bertie, Bass Islands (*see* Sequence 6) or Oriskany formations is usually marked by several metres of relief (Uyeno et al. 1982) and the development of paleokarst (Kobluk et al. 1977), suggesting a significant depositional hiatus. Weathered fractures in the upper part of the Bertie Formation are commonly filled with coarse sand and well-rounded pebbles and cobbles from Devonian clastic units. Lenses of well-rounded pebbles and cobbles are also reported at this contact (Derry, Michener, Booth and Wahl and OGS 1989c).

The Bois Blanc Formation is overlain by the Amherstburg Formation (of the Detroit River Group) west of Hagersville, and by the Onondaga Formation (Edgecliffe Member) east of Hagersville (Uyeno et al. 1982). Due to poor exposure and undiagnostic subsurface data, the Bois Blanc–Amherstburg contact is commonly difficult to determine and is therefore somewhat arbitrarily defined; this results in variable total thicknesses being reported for the Bois Blanc Formation.

Overlying the Bois Blanc Formation is a complex succession of limestones and dolostones of Early to Middle Devonian age. There is some uncertainty associated with placement of the Lower–Middle Devonian boundary within this succession (e.g., Sanford 1968; Winder and Sanford 1972; Uyeno et al. 1982; Derry, Michener, Booth and Wahl and OGS 1989a). Recent conodont biostratigraphic work on adjacent strata in Ohio (Sparling 1988) indicates that this boundary occurs in the lower parts of the Amherstburg (Detroit River Group) and Onondaga formations (Chart D, Map 2582, map case; *see* Figure 20.35).

In the Niagara Peninsula area, the Bois Blanc Formation is overlain (possibly disconformably) by cherty limestones of the Onondaga Formation (Telford and Tarrant 1975a, 1975b). The Onondaga Limestone of western New York State has 4 members, 3 of which (the Edgecliffe, Clarence and Moorehouse members) extend into Ontario (*see* Figure 20.35). In Ontario, the distinctive characteristics of these members merge, westward (Telford and Tarrant 1975a), as they grade laterally into the Detroit River Group of the Michigan Basin.

The basal Edgecliffe Member and the uppermost Moorehouse Member are both typically biostromal limestones with chert nodules (*see* Table 20.7; Telford and Tarrant 1975a, 1975b; Uyeno et al. 1982). The Edgecliffe beds often surround and blanket biohermal mounds constructed mainly of the colonial rugose coral *Acinophyllum* (Telford and Tarrant 1975b; Uyeno et al. 1982). The Clarence Member is a very cherty, relatively unfossiliferous limestone. Bioherms of the Edgecliffe Member usually extend up into the Clarence Member, causing deformed and domed beds. The contact with the overlying Moorehouse Member is gradational. The contact between the top of the Moorehouse Member and the overlying Dundee Formation is sharp and well exposed on the Niagara Peninsula (Uyeno et al. 1982; Derry, Michener, Booth and Wahl and OGS 1989c).

The Detroit River Group (*see* Figure 20.35; Chart D, Map 2582, map case), overlying the Bois Blanc Formation in southwestern Ontario, is one of the least understood Devonian units in Ontario. Poor exposure and a lack of biostratigraphic control have hindered development of subdivisions and correlation. Three formations, the Sylvania, Amherstburg and Lucas (*see* Figure 20.35 and Table 20.7), are included in this group (Uyeno et al. 1982; Derry, Michener, Booth and Wahl and OGS 1989a). Status of the Sylvania Formation is equivocal. Some workers suggest a formational status, whereas others suggest it is a member within the Amherstburg Formation (e.g., Landes 1951; Sanford 1968; Fagerstrom 1971). The Sylvania Formation has also informally been referred to as the "Sylvania Sandstone".

The Sylvania Formation is an orthoquartzitic sandstone restricted in southwestern Ontario to the subsurface between Windsor and Sarnia (Uyeno et al. 1982). This Michigan Basin unit represents reworking of older craton-derived clastic sediments. It has been interpreted as a strand line or eolian deposit associated with a minor transgression that occurred in the early part of Eifelian time (Uyeno et al. 1982).

The Amherstburg Formation is a relatively thick (*see* Table 20.7) carbonate unit which overlies and, in part, grades laterally into the Sylvania Formation. In the Ingersol–Beachville area, the unit is generally a bioclastic, cherty and bituminous limestone (Derry, Michener, Booth and Wahl and OGS 1989c), whereas in the western Lake

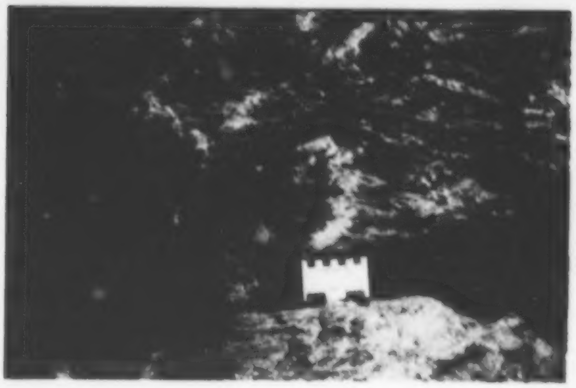

**Figure 20.36.** Stromatoporoid-rich outcrop of the Formosa Reef Limestone, Amherstburg Formation, in roadcut north of Formosa (photograph *courtesy of* B. Feenstra).

Erie–Detroit River area, it is typically a more massive-bedded, bituminous dolostone with oolitic, bioclastic and coralgal zones (Uyeno et al. 1982).

The Amherstburg Formation includes a localized, but well-developed, reef complex, the Formosa Reef Limestone (Fagerstrom 1961; Klein 1980). This complex is up to 15 m thick and extends over approximately 390 km² in Bruce and Huron counties, with its type section located 4 km north of the village of Formosa (Fagerstrom 1961). Lithologically, the reefs are grey, fine-grained, high purity limestones. Stromatoporoids are the principal frame- builders (Figure 20.36), with subordinate tabulate and rugose corals. Other fossils include brachiopods, molluscs and trilobites (Best 1953; Klein 1980). Klein (1980) identified multiple stages of reef growth within the complex. The Formosa Reef complex appears to represent a shallower water environment of deposition than the well-bedded, more basinal, bituminous limestones which characterize the remainder of the Amherstburg Formation.

Conformably overlying the Amherstburg Formation is the Lucas Formation. In the central part of the Michigan Basin, this unit consists of limestones and dolostones alternating with anhydritic dolostones, anhydrite and halite (Sanford 1968). Along the basin margins, for example in the Algonquin Arch region of southwestern Ontario, the Lucas Formation consists of high purity limestones and bituminous and cherty dolostones (Figure 20.37; Sanford 1968; Uyeno et al. 1982). Drill cores from southwestern Ontario have shown the Lucas to have thin anhydrite-gypsum beds, partings and blebs (e.g., Johnson et al. 1985). Brecciation within the Lucas Formation suggests there has been removal of evaporites by dissolution. A "needle-like" moldic porosity, after evaporite minerals, is also commonly present.

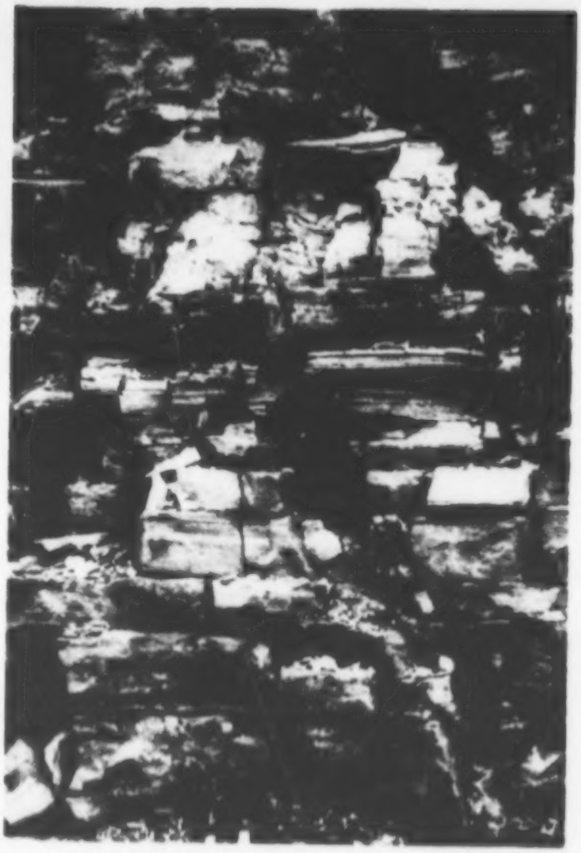

**Figure 20.37.** Lucas Formation limestone in the St. Marys Cement Quarry, at St. Marys. Its upper contact with the more fossiliferous, darker grey limestone of the Dundee Formation is indicated (white bar) near the top of this photograph (photograph *courtesy of* R. Kelly).

At the top of the Lucas Formation, and at some localities constituting the entire formation, is the Anderdon Member (*see* Figure 20.35), a micritic high calcium limestone alternating with coarser bioclastic limestone. A sandy limestone lithofacies occurs within the unit and has been termed the "Columbus Sands" by the oil and gas industry. This name has been applied even though these beds have not been proven to be stratigraphically equivalent to their namesake, the Columbus Limestone of Ohio (Sanford 1968; Birchard 1990; Birchard and Risk 1990). This sandy carbonate lithofacies is widespread and commonly seen in well cuttings from Kent and Essex counties, and as far east as Ingersol.

The Lucas Formation thins toward the margins of the Michigan Basin and pinches out in the Port Dover area on the Niagara Peninsula, placing the underlying Amherstburg Formation in direct contact with the overlying Dundee Formation (Telford and Hamblin 1980). Whether this contact represents a regional unconformity is unclear. Uyeno et al. (1982) presented the following 3 arguments in support of an unconformable contact: 1) the Lucas–Dundee contact is sharp (*see* Figure 20.37) and undulatory, with the lower 5 to 10 cm of the Dundee Formation displaying reworked Lucas debris; 2) the sandy interval at the top of the Anderdon Member implies a period of emergence, sand deposition and subsequent submergence (Summerson and Swann 1970); and 3) there is a major change in the conodont fauna across this contact. It is noteworthy that a hiatus also exists at the contact between approximately equivalent strata on the Hudson Platform (the Moose River and overlying Murray

Island formations), supporting the regional significance of this event (Sanford and Norris 1975; Uyeno et al. 1982).

The richly fossiliferous, micritic limestones of the Dundee Formation subcrop beneath Quaternary sediments in a broad, northwest-trending belt in southwestern Ontario, extending from central Lake Erie to Lake Huron, and also underlie parts of the Windsor–Essex area and Pelee Island in Lake Erie (Map 2544, map case). The Dundee Formation is correlative, in part, with the Delaware Formation of Ohio and the upper part of the Onondaga Formation in New York (Uyeno et al. 1982). The Tioga Ash Bed B, which separates the Moorehouse and Seneca members of the Onondaga Formation in New York and which is widely used as a marker horizon in the Appalachian Basin (Rickard 1984), may be present in the lower part of the Dundee Formation, but confirmation is still required (Sanford 1968; Uyeno et al. 1982; Johnson et al. 1989). For the most part, the contacts of the Dundee Formation with the underlying Lucas Formation and the overlying Hamilton Group (*see* Sequence 8) are sharp and erosional (Hamilton and Coniglio 1990).

The Dundee Formation is thickest (120 m) in the Saginaw Bay area of Michigan (Sanford 1968), but is considerably thinner in southwestern Ontario (*see* Table 20.7). Typical thicknesses of 35 to 45 m were encountered during recent regional deep stratigraphic drilling by the OGS (*see* Figure 20.5; Johnson et al. 1985).

The Dundee Formation in Ontario consists mainly of fossiliferous limestones with bituminous partings and chert nodules. On Pelee Island, the unit is more massive, lighter coloured and bioclastic in nature. On the Niagara Peninsula, the Dundee Formation contains prominent horizons of black bituminous limestone. Bituminous material is commonly found in vugs and is concentrated in the porous frameworks of tabulate corals (Uyeno et al. 1982; Derry, Michener, Booth and Wahl and OGS 1989c); commonly, samples from this unit have a distinctly petroliferous odour when freshly broken.

The Dundee Formation is characterized by an abundant and diverse brachiopod fauna, in addition to molluscs, bryozoans and trilobites. Exposures of the Dundee Formation on Pelee Island contain the most diverse fauna of this unit in southwestern Ontario, displaying, in addition to those fauna mentioned above, solitary and colonial rugose corals, and tabulate corals (Uyeno et al. 1982).

The Dundee Formation was deposited in lagoonal, open shelf, and deep water environments, with its lithological character and distribution being largely controlled by the structural highs (Findlay and Algonquin arches) separating the 2 depositional basins (Birchard 1990). The presence of a number of laterally persistent "firmgrounds" (early, partly cemented surfaces) in the unit suggests that deposition was periodically interrupted by relative sea level fluctuations (Birchard 1990; Birchard and Risk 1990).

In most of southern Ontario, the top of the Dundee Formation marks the close of Sequence 7 deposition. However, between approximately Port Stanley and Long Point on Lake Erie, a wedge of Appalachian Basin clastic rocks, the Marcellus Formation, extends inland approximately as far as Aylmer and St. Thomas, pinching out between the Dundee Formation and overlying Hamilton Group (*see* Sequence 8) rocks (Map 2544; Chart D, Map 2582, map case).

The Marcellus Formation consists mainly of black, organic-rich shales, with minor, thin, impure carbonate interbeds. The organic content, mostly in the form of kerogen (with lesser bitumen), is mainly concentrated within 2 organic-rich zones, separated by a carbonate-rich interval. These upper and lower organic-rich zones average 7.38 and 4.14% in total organic carbon content, respectively (Brown 1985; Johnson et al. 1989). The organic properties and oil shale potential of the Marcellus Formation are discussed in the Economic Geology section.

The lower contact of the Marcellus Formation with the Dundee Formation is apparently conformable, with thin organic-rich interbeds occurring in the uppermost metre of the Dundee Formation (Telford 1984; Johnson et al. 1989). The Marcellus Formation's upper contact with the Bell Formation shales (Hamilton Group; Sequence 8) is sharp and disconformable but lacks conclusive evidence of erosion. Locally, minor, black, organic-rich shale interbeds persist up into the Bell Formation (Johnson et al. 1989).

## NORTHERN ONTARIO

In northern Ontario, Sequence 7 deposits, consisting of carbonates, evaporites and lesser craton-derived clastic rocks, are lithologically and faunally similar to the equivalent succession of southern Ontario, which suggests probable direct seaway connections between the Hudson Platform and the St. Lawrence Platform during this time (Figure 8, Map 2582, map case; Telford 1988). However, no Devonian strata are known in the Lake Timiskaming outlier (Russell 1984) and no intervening outliers have been found to confirm such connections.

During the Late Silurian (*see* Sequence 6), a major northwest-trending Precambrian horst, the Central Hudson Bay Arch, was reactivated, splitting the Hudson Bay Basin into eastern and western subbasins (*see* Figure 20.30; Sanford 1987; Thorpe 1988; Sanford and Grant 1990). Uplift of the horst and a possible eustatic sea level fall (in the Early Devonian) affected the sediment distribution and thickness of the uppermost Sequence 6 units and some Sequence 7 units (i.e., the upper Kenogami River, Stooping River and lower Kwataboahegan formations; Roksandic 1987; Sanford 1987; Thorpe 1988). During the early Middle Devonian, a transgression and change in the tectonic regime of the area resulted in subsidence and burial of the central high and a rejoining of the 2 subbasins for the remainder of Sequence 7 time (Roksandic 1987).

The Sextant Formation, representing initial Sequence 7 deposition on the Hudson Platform, is geographically restricted to the southern margin of the Moose River Basin. This unit consists of a variety of clastic rock types, but is predominantly a reddish arkosic sandstone (Sanford and Norris 1975; Telford 1988). The Sextant Formation contains one of the earliest Devonian land plant assemblages, with both a diverse and well-preserved flora (Lemon 1953; McGregor et al. 1970; McGregor and Camfield 1976;

Hueber 1983). The Sextant Formation is generally regarded as nonmarine (Sanford et al. 1968; Sanford and Norris 1975); however, marginal marine interpretations have also been proposed (Stoakes 1978).

Overlying and laterally equivalent, in part, to the Sextant Formation is the Stooping River Formation, a cherty limestone with lesser dolomitic limestone and dolostone (Sanford et al. 1968; Telford 1988). Many of the Stooping River beds are richly fossiliferous (Sanford and Norris 1975). The fauna in this unit suggest a possible correlation with the Bois Blanc and lower Onondaga formations of southern Ontario (Chart D, Map 2582, map case; Telford 1988). In addition to its vertical and lateral gradational and conformable contact with the Sextant Formation, the Stooping River Formation unconformably overlies Precambrian, Ordovician, Silurian and Lower Devonian strata in the Moose River Basin (Sanford and Norris 1975). Most of the Stooping River Formation was probably deposited in intertidal to shallow subtidal environments and represents a return to fully marine conditions (Telford 1988).

Overlying the Stooping River Formation is the Kwataboahegan Formation, an extremely fossiliferous (Figure 20.38) succession of biostromal and biohermal limestones (Sanford et al. 1968). Thickness variation of this unit (*see* Table 20.7) is probably due to biohermal buildups. The best outcrops of the Kwataboahegan Formation occur at Coral Rapids, in the Moose River Basin, where the limestone outcrops rise 15 m above river level for several kilometres.

Carbonate buildups within the Kwataboahegan Formation appear to be associated with topographic highs on the Precambrian basement surface and are interpreted as shelf-edge reefal deposits (Telford 1988). Thinner-bedded, locally bituminous limestones and minor dolostones occur between the bioherms and are interpreted as interreefal deposits (Sanford and Norris 1975). Conodont evidence indicates similar ages for the Stooping River Formation and lower part of the Kwataboahegan Formation, suggesting that parts of these units were deposited contemporaneously (Telford 1988).

Overlying the Kwataboahegan Formation are the evaporites and evaporitic, brecciated carbonates of the Moose River Formation. Spectacular outcrops of almost pure, white gypsum (Figure 20.39) occur along the Moose, Abitibi and Cheepash rivers in the Moose River Basin (Bezys 1990). Anhydrite has only been observed in this formation in the subsurface (Bezys 1990). The distribution of gypsum in this unit in the Moose River Basin appears to have been controlled by north to northwest-trending structural highs (i.e., Pivabiska Ridge and the Moose River and Grand Rapids highs; *see* Figure 20.6; Bezys 1990).

The Moose River Formation in the Hudson Bay Basin is primarily known from well records, and contains a significant thickness of halite (Sanford and Norris 1975; Sanford 1987; Thorpe 1988). Although no halite deposits have been preserved in this formation in Moose River Basin, chaotic breccias in the Moose River Formation suggest that halite had accumulated in association with gypsum but was removed by subsequent dissolution (Sanford and Norris 1975). Brecciation is very extensive in the Moose River Formation and is characterized by exotic rock fragments and a distinctive green mudstone matrix (Bezys 1990). The

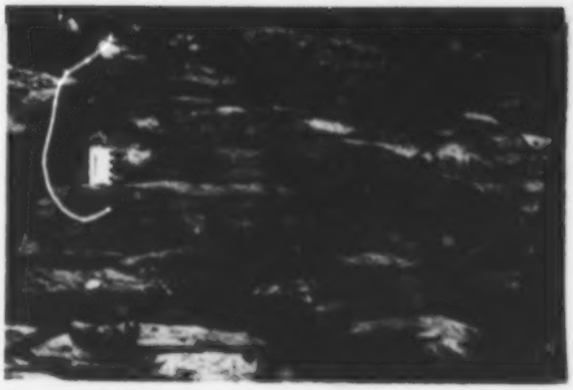

**Figure 20.38.** Fossiliferous (corals and stromatoporoids) limestone of the Kwataboahegan Formation, exposed on a solution-enhanced joint surface in the Labelle Quarry, at Moosonee. (Photo: DA91–31–17)

**Figure 20.39.** Massive gypsum beds of the Moose River Formation along the Cheepash River, James Bay Lowland. (Photo: RB88–13–3)

thickness of this formation is highly variable in the Moose River Basin (Telford 1988), which suggests that dissolution affected local unit thicknesses.

The Moose River Formation represents a drop in relative sea level which resulted in the precipitation of evaporites in a restricted shallow marine environment, which, by the end of Moose River time, was a large sabkha (Thorpe 1988; Telford 1988).

The Murray Island Formation consists of a thin interval of fossiliferous limestones (Sanford and Norris 1975). The brachiopod fauna in this unit is correlative with that of the upper Onondaga and Dundee formations of southern Ontario (Norris 1986).

Separating the Murray Island Formation from the underlying Moose River Formation is a regional disconformity of short duration (Chart D, Map 2582, map case; Norris 1986; Sanford and Grant 1990). Norris (1986) reported reworked clasts of Moose River sediments in the basal beds of the Murray Island Formation. Drill cores through this interval (e.g., OGS Onakawana B and Schlievert Lake drill holes) show little evidence of extended exposure. The contact relationship between these units is commonly obscured by brecciation in the lower part of the Murray Island Formation (Russell et al. 1985; Sanderson and Telford 1985; Bezys 1989).

The upper contact of the Murray Island Formation with overlying shales and carbonates of the Williams Island Formation (*see* Sequence 8) is sharp and probably disconformable (Sanford and Norris 1975). The variable thickness of the Murray Island Formation has been attributed to erosion that occurred during this hiatus, prior to deposition of the Williams Island Formation (Telford 1988).

## Economic Geology

The first recorded evidence of oil in southern Ontario, dated from the early 19th century, described seeps from shallow reservoirs of Devonian age. The first oil well in North America was completed in 1858 at Oil Springs in southwestern Ontario, and discovered crude oil in the Dundee Formation (Carter 1990). Since then, 25 to 40% of Ontario's oil production (6.65 million $m^3$ to end 1989) has come from Devonian traps (Carter 1990). In 1989, oil production from Devonian strata totalled 32 740 $m^3$, accounting for 13.5% of the total Ontario production (Carter 1990). Production has come from 25 pools of which 7 are still in production (Carter 1989). Some 88% of this total Devonian production has come from 3 reservoirs: the Oil Springs, Petrolia and Rodney pools. To date, oil production from Devonian units has been largely restricted to the Dundee and Lucas formations. Minor amounts of oil and gas have been found in the overlying Hamilton Group (Sequence 8) carbonate beds, although this unit primarily acts as a top seal for the Dundee–Lucas reservoirs.

Production comes from 2 types of traps within the Devonian: 1) high porosity zones in the sandy facies of the Anderdon Member (often termed "Columbus" or "oil sand" by the oil industry), particularly at the interfingering of this facies with the remainder of the Anderdon Member; and 2), carbonate traps in the Rockport Quarry (*see* Sequence 8), Dundee and Lucas formations. Both of these traps are associated with fractures and structural highs, although they have limited closure (Bailey and Cochrane 1985). As fractures form an interconnected network, initial flow rates can be very high (e.g., 7500 barrels (1200 $m^3$) per day from wells at Oil Springs) from this type of trap (Bailey and Cochrane 1985).

All of the Sequence 7 carbonate units are quarried for a variety of limestone products including aggregate, building stone, armour stone, lime and cement (Bezys and Johnson 1988). The silica content of some units restrict their use as concrete aggregate. The Anderdon Member limestone of the Lucas Formation, extracted from several quarries in Amherstburg and the Woodstock to Ingersol areas, is of high purity and used by the steel, cement and chemical industries (Derry, Michener, Booth and Wahl and OGS 1989c).

Sink hole development in the Middle Devonian strata of the Windsor area has attracted much attention in recent years. Studies indicate that sink hole development may be related to solution mining of salt beds of the underlying Salina Formation (namely, the Salina B, D and F units), lying a minimum of 285 m below the ground surface (*see* Economic Geology of Sequence 6; Nieto et al. 1983). Nieto et al. (1983) proposed that when the initial collapse propagated from the Salina Formation up to the level of the Sylvania Formation (a soft, friable sandstone unit lying 135 m below the surface), the latter unit failed and its uncon-solidated grains then filtered down the joints and fractured mass, thereby facilitating propagation of the collapse to surface.

During the mid–1980s, the OGS conducted a study of the Marcellus Formation to determine its organic properties and potential as an oil shale (Barker 1985; Johnson 1985; Johnson et al. 1989). The study found that the oil content (in a kerogen equivalent form) of the Canadian part of the unit is about $5.8 \times 10^9$ barrels and that the part of this oil-equivalent kerogen that falls below the land area of Ontario is about $3.7 \times 10^9$ barrels. Although the volume of oil contained within this unit is quite staggering, its development potential is limited. Most of the unit is buried below overlying Paleozoic strata, and even the most accessible parts of this unit are buried below over 80 m of Quaternary drift. Based on the thickness of the organic zones, a stripping ratio of around 50:1 is indicated, making extraction both expensive and difficult based on available technology. The nearby Upper Devonian Kettle Point Formation (*see* Sequence 8) and Upper Ordovician Collingwood Member (*see* Sequence 2) oil shales have greater thicknesses, underlie larger areas and can be found under thinner drift cover, thus making them more attractive alternatives (Russell and Telford 1983; Russell 1985).

The gypsum deposits of the Moose River Formation (*see* Figure 20.39) have attracted considerable interest since first noted by geologists over 100 years ago (Bezys 1990). The remote location of these deposits has limited their development; however, the exceptional purity of this gypsum may ultimately lead to development of these deposits for specialty filler markets (Hains Technology Associates and Bezys 1990).

# DEPOSITIONAL SEQUENCE 8: LATE MIDDLE DEVONIAN (GIVETIAN) TO EARLY MISSISSIPPIAN (TOURNAISIAN)

## Overview

Sequence 8, characterized by orogen-derived clastic sediments with minor carbonates, was deposited in a seaway which covered most of eastern North America during the main phase of the Acadian Orogeny. These sediments unconformably overlie the widespread carbonates of Sequence 7 and are the youngest Paleozoic strata preserved in Ontario. Sequence 8 is unconformably overlain by Quaternary sediments in southern Ontario and Mesozoic age sediments of Sequence 9 in the Hudson Platform.

Commencing during the late Middle Devonian, the Acadian Orogeny, along the eastern margin of the North American craton, produced a major change in the sedimentation patterns within the Appalachian and Michigan basins. Thick, nonmarine, red bed clastic sediments of the Catskill Delta were deposited in the southern part of the Appalachian Basin during Givetian to Famennian time, reflecting intensive orogenic movement along the adjacent Appalachian structural front (Figure 9, Map 2582, map case; Oliver et al. 1968; Colton 1970; Beardsley and Cable 1983). These orogen-derived clastic sediments intertongue with marine shales and minor carbonate sediments in the northern Appalachian Basin and adjacent regions of the Michigan Basin (Figure 9, Map 2582, map case; Cohee 1947; Cohee et al. 1951; Sanford 1968).

Sequence 8 sediments were deposited in a broad, shallow, inland seaway, known as the Chattanooga Sea (Figure 20.40), which covered much of the eastern United States and southern Ontario (Janka and Dennison 1980). This seaway extended into the Central St. Lawrence Platform (based on Givetian age fossils in sedimentary rock fragments in the Sainte-Helene Island breccia at Montreal, Boucot et al. 1986) and was also connected with the Hudson Platform, where shales and limestones were deposited in the Moose River and Hudson Bay basins (Figure 9 and Chart D, Map 2582, map case; Sanford and Norris 1975).

Over most of eastern North America, much of Sequence 8 sedimentation is characterized by organic-rich black shales (Figure 20.41). The conditions necessary to create these widespread deposits may have resulted from a combination of high biological productivity in this equatorial sea and poor water circulation created by its semi-restricted paleogeographic setting (e.g., Demaison and Moore 1980). Even though this seaway was open to the south, it would have been bounded on the east by the Acadian Orogen and confined in the north and west by Precambrian and Lower Paleozoic rocks of the stable continental craton.

The greatest accumulation of Sequence 8 shales took place in the Appalachian Basin part of the Chattanooga Sea (*see* Figures 20.40 and 20.41), adjacent to the Acadian Orogen. Shale thickness varies from several thousand metres in West Virginia, in the east-central Appalachian Basin, to 100 to 300 m in the Illinois and Michigan basins (Janka and Dennison 1980). The depositional centres of the latter 2 basins were separated from the larger Appalachian Basin by the Cincinnati, Findlay and Algonquin arches (*see* Figure 20.2). Thus, although these arches and other structural highs, such as the Nashville Dome, may not have completely impeded water circulation amongst the basins, they did have a significant effect on the regional sediment depositional patterns.

The shale-dominated Sequence 8 succession of the Moose River Basin is lithologically similar to that of southwestern Ontario (Norris 1986; Telford 1988). This suggests that a seaway connected these 2 areas and that these clastic sediments were derived from a common source, namely, the highlands of the Acadian Orogeny (Sanford 1987; Telford 1988). In the Hudson Bay Basin, on the other hand, local cratonic sources (i.e., the arches that bordered the basin on the north, east and west) are indicated by coarse, immature, shallow water clastic sediments which constitute approximately equivalent strata there (Figure 9, Map 2582, map case; Sanford 1987). The lithological differences between Upper Devonian strata in the Hudson Bay Basin and the Moose River Basin suggests that the Cape Henrietta Maria Arch may have acted as a barrier to deposition at this time (Telford 1988).

The seaway connection between the Hudson and Interior platforms (via the Williston Basin), which was apparently severed in the Late Silurian (*see* Sequence 6), was possibly re-established in the late Middle Devonian, as indicated by lithologic and faunal similarities of Givetian age rock units in these 2 regions (Sanford 1987).

The detailed stratigraphy of Sequence 8 is described in the following section and illustrated in Figure 20.42. Lithologic details of the units are presented in Table 20.8. The existing distribution of these units is shown on Maps 2541 to 2544 (map case), and that of the sequence, as a whole, is shown on Maps 2575 to 2578 (map case). The pre-erosional maximum and present distributions of the sequence and its constituent sediment types are shown on Figure 9, Map 2582 (map case).

## Geology

### SOUTHERN ONTARIO

The base of Sequence 8 is marked by a brief but significant break in deposition, which separates this shale-dominated succession from the underlying carbonate-dominated strata of Sequence 7 (Uyeno et al. 1982). In most of southern Ontario, this contact is distinct between the shales of the Hamilton Group (Bell Formation) and the underlying Dundee Formation carbonates (Chart D, Map 2582, map case). However, along the north shore of central Lake Erie, where the uppermost Sequence 7 strata consist of shales of

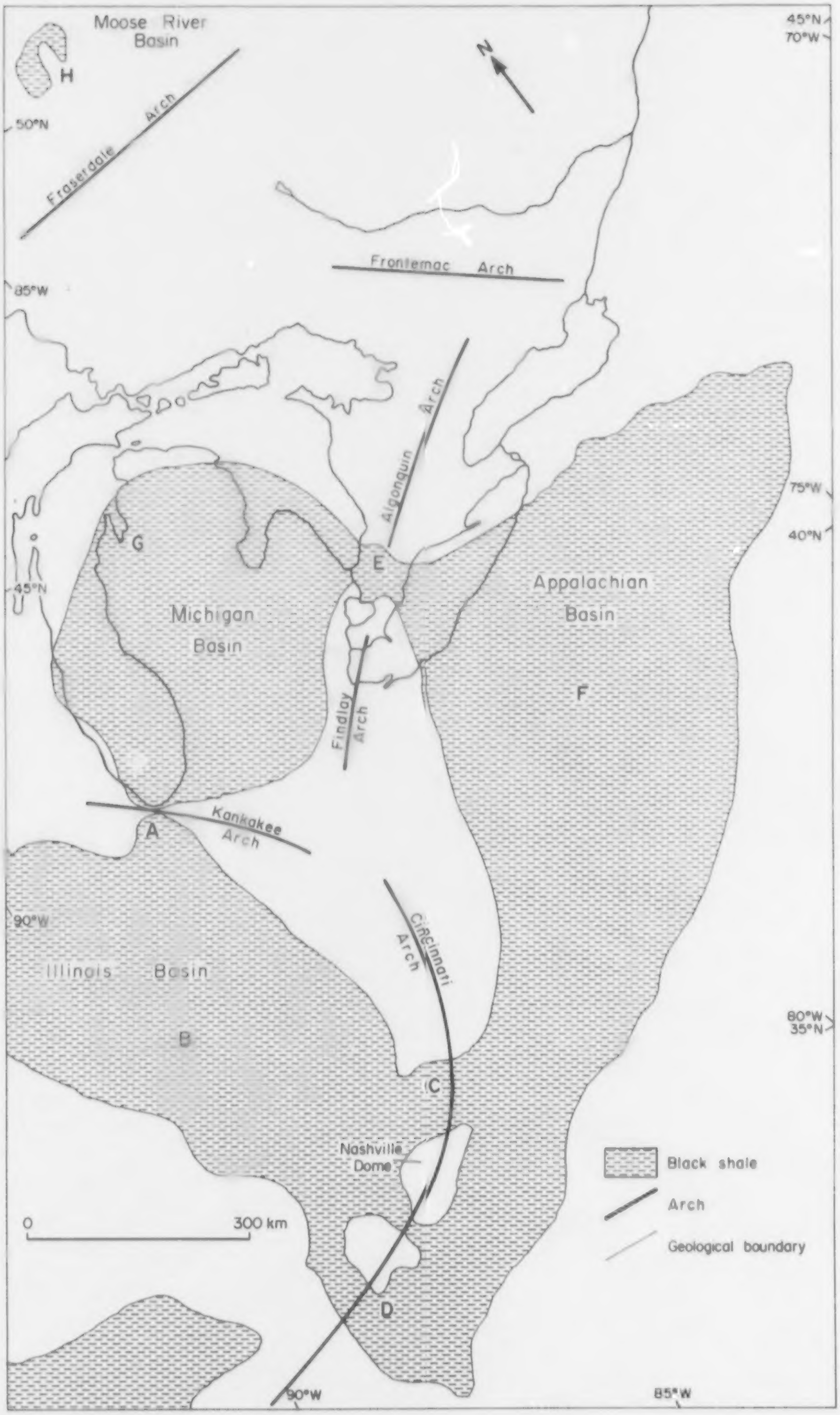
Moose River Basin
H
Fraserdale Arch
50°N
N
45°N
70°W
Frontenac Arch
85°W
Algonquin Arch
75°W
40°N
G
E
45°N
Michigan Basin
Appalachian Basin
F
Findlay Arch
A
Kankakee Arch
90°W
Cincinnati Arch
Illinois Basin
B
80°W
35°N
C
Nashville Dome
Black shale
Arch
Geological boundary
0
300 km
D
90°W
85°W

**Figure 20.40.** Present distribution of Upper Devonian black shales in eastern North America, indicating extent of the Late Devonian Chattanooga Sea. Letters A to H denote locations of sections shown in Figure 20.41 (*modified from* Russell 1985).

the Marcellus Formation, this contact is less distinct (Johnson et al. 1989).

The Hamilton Group in southwestern Ontario consists mostly of mudstone and shale with thin, impure, laterally persistent carbonate horizons which have been accorded formation status (Uyeno et al. 1982). The formations constituting the Hamilton Group are, in ascending order, the Bell, Rockport Quarry, Arkona, Hungry Hollow, Widder and Ipperwash formations (*see* Figure 20.42 and Table 20.8).

The present distribution of the Hamilton Group in the southwestern Ontario peninsula is limited to an area between Lake Erie and Lake Huron (Map 2544, map case), largely within the Chatham Sag area (*see* Figure 20.2; Uyeno et al. 1982). Although the formations are mostly covered by Quaternary material, bedrock topography (Karrow 1973) indicates that the erosional edge of the Hamilton Group forms a distinct escarpment.

The Hamilton Group in southwestern Ontario is considered to be of early Givetian age (Chart D, Map 2582, map case; Uyeno et al. 1982). Biostratigraphic control for the lower (mostly subsurface) units (Bell to lower Arkona formations) is poor; these formations may be as old as late Eifelian (Johnson et al. 1989). The Hamilton Group in Ontario correlates with the middle part of the more clastic-rich Hamilton Group of New York State and with the more carbonate-rich Traverse Group of Michigan (Uyeno et al. 1982), and thus occupies a lithologic transition zone between the Appalachian and Michigan basins (Figure 9, Map 2582, map case).

The basal Bell Formation is a soft shale, with limestone and organic-rich shale interbeds (Uyeno et al. 1982; Johnson et al. 1989). The unit is not exposed in southwestern Ontario and OGS drill cores show it to be a homogeneous unit, rich in shelly debris (Johnson et al. 1985).

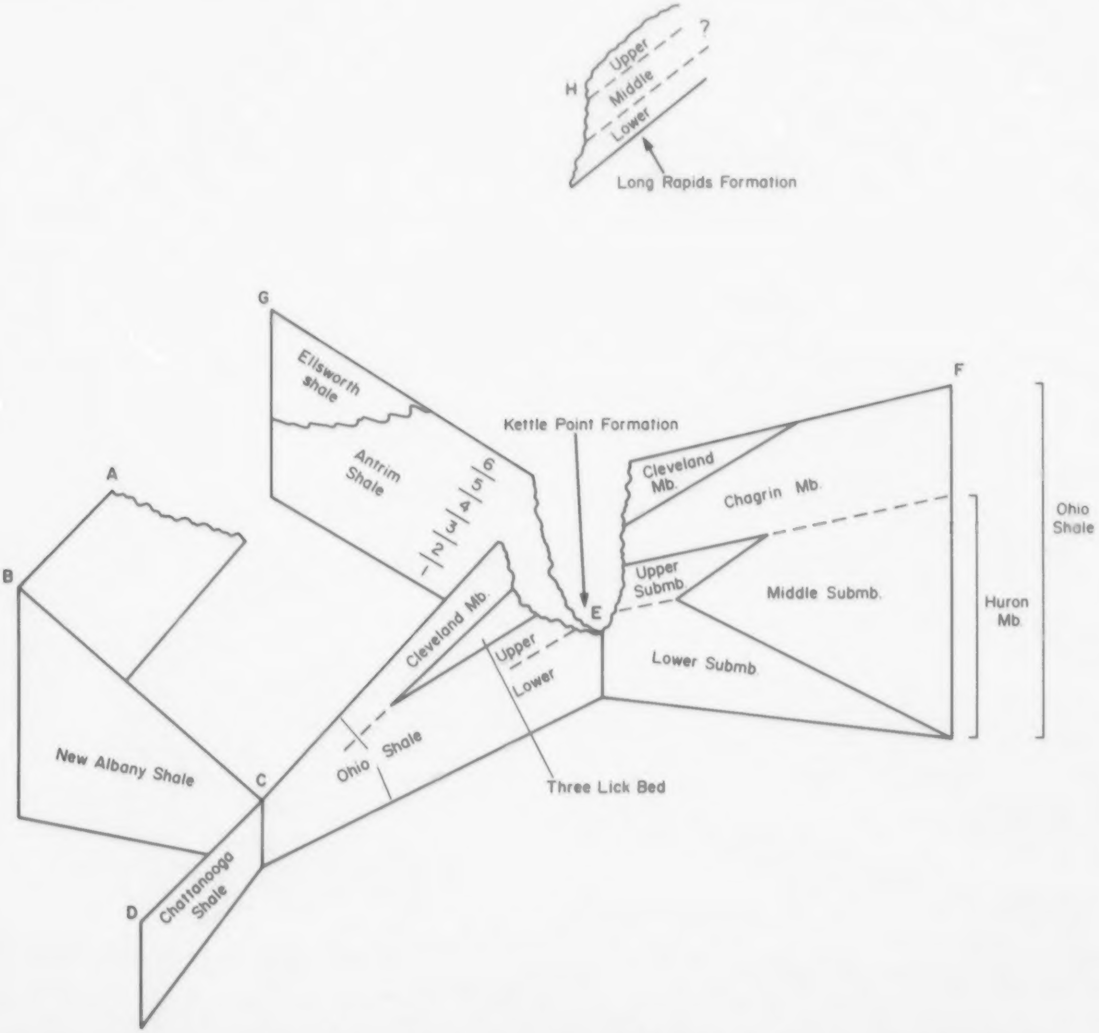
H
Upper
Middle
Lower
?
Long Rapids Formation
G
Ellsworth Shale
Antrim Shale
1
2
3
4
5
6
Kettle Point Formation
F
Cleveland Mb.
Chagrin Mb.
Upper Submb.
Middle Submb.
Lower Submb.
Ohio Shale
Huron Mb.
A
B
E
Cleveland Mb.
Upper
Lower
Ohio Shale
Three Lick Bed
New Albany Shale
C
D
Chattanooga Shale

**Figure 20.41.** Diagrammatic representation of stratigraphic relationships of the Upper Devonian black shales of eastern North America. Letters A to H represent locations shown in Figure 20.40 (*modified from* Russell 1985).

The overlying Rockport Quarry Formation consists of very fine-grained limestone (Sanford 1968) with sharp upper and lower contacts. The unit becomes increasingly shaly to the southeast and appears to grade laterally into the Arkona Formation.

The Arkona Formation consists of relatively unfossiliferous, blue-grey shales with local concentrations of brachiopods (Uyeno et al. 1982). The uppermost part of this unit is well exposed along the Ausable River near the villages of Arkona and Thedford, approximately 50 km west of London (Telford and Johnson 1984).

The overlying Hungry Hollow Formation is one of the thinnest Paleozoic formations identified in Ontario. The unit is less than 2 m thick, but has remarkable lateral persistence and can be subdivided into 7 (Uyeno et al. 1982) or 9 (Mitchell 1967; Landing and Brett 1987) biostratigraphic and/or lithostratigraphic horizons. This unit is also well exposed along the Ausable River near Arkona.

Conformably overlying the Hungry Hollow Formation is the Widder Formation, which consists of calcareous shales with interbedded argillaceous, crinoidal and nodular limestones (Uyeno et al. 1982). The unit is richly fossiliferous in brachiopods, bivalves, cephalopods and trilobites (Uyeno et al. 1982).

The uppermost unit of the Hamilton Group is the Ipperwash Formation, a richly fossiliferous, bioclastic limestone,

| | **Northern Ontario** | **Southern Ontario** | |
|---|---|---|---|
| **Series** | **Hudson Bay and Moose River basins** | **Appalachian and Michigan basins** | |
| Upper Devonian | | Port Lambton Gp.* | Sunbury Fm.* |
| | | | Berea Fm.* |
| | | | Bedford Fm.* |
| | Long Rapids Fm. | Kettle Point Fm. | |
| Middle Devonian | Williams Island Fm. | Hamilton Gp. | Ipperwash Fm. |
| | | | Widder Fm. |
| | | | Hungry Hollow Fm. |
| | | | Arkona Fm. |
| | | | Rockport Quarry Fm. |
| | | | Bell Fm. |

**Figure 20.42.** Stratigraphic units of Sequence 8. *May, in part, be Mississippian in age; not present in Appalachian Basin in Ontario.

**Table 20.8.** Middle Devonian (Givetian) to Early Mississippian (Tournaisian) units in Ontario—Depositional Sequence 8 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Southern Ontario** | | | | | | |
| **Sunbury Formation (Port Lambton Group)** | Lambton County area | Famenian and/or Tournaisian(?) | Marine environment with stratified water column | 20 m (max.) | Black, organic-rich shales.<br>Lower contact is sharp and disconformable with Berea Fm. | Uyeno et al. 1982;<br>Johnson et al. 1985 |
| **Berea Formation (Port Lambton Group)** | Lambton County area | Famennian and/or Tournaisian(?) | Shallow marine conditions | 60 m (max.) | Grey, fine- to medium-grained sandstones interbedded with grey shale and siltstone.<br>Sandstone channels ("shoe-strings") cut down into the Kettle Point Fm.<br>Lower contact is gradational with Bedford Fm. and disconformable with Kettle Point Fm. | Sanford 1968;<br>Uyeno et al. 1982 |
| **Bedford Formation (Port Lambton Group)** | Lambton County area | Famennian and/or Tournaisian(?) | Shallow marine conditions | ~30 m | Light grey, soft, fissile shales; homogeneous and unfossiliferous.<br>Upper part of unit has silty and sandy interbeds.<br>Lower contact is sharp, erosional and disconformable with Kettle Point Fm. | Johnson et al. 1985;<br>Sanford 1968 |
| **Kettle Point Formation** | Chatham and Sarnia areas | Frasnian to Famennian | Marine environment with stratified water column | 30 to 75 m | Black, siliciclastic, organic-rich shale and siltstone with minor green-grey, organic-poor shale and siltstone interbeds. Organic-rich intervals (3 to 15% organic carbon) are laminated. Organic-poor (<2 % organic carbon) interbeds are bioturbated. Large (up to 1.2 m in diameter) calcite concretions or "kettles" in lower one-third of Kettle Point Fm. Abundant tasmanites, pyritized radiolaria, sponge spicules, conodonts, plant remains. Rare lingulid brachiopods and fish remains. Basal lag bed.<br>Lower contact is sharp and disconformable with Ipperwash Fm. | Russell 1985;<br>Uyeno et al. 1982;<br>Winder and Sanford 1972;<br>Sanford 1968;<br>Armstrong 1986;<br>Russell and Barker 1983;<br>Winder 1966 |
| **Ipperwash Formation (Hamilton Group)** | Area between Lake Erie and Lake Huron | Givetian | Shallow marine | 2 to 13 m | Grey-brown, medium- to coarse-grained, bioclastic limestone; richly fossiliferous; minor chert.<br>Fossils include crinoids, bryozoans, corals, bivalves. Large burrows (15 cm in diameter) in the Ipperwash Beach area.<br>Lower contact is sharp with Widder Fm. | Uyeno et al. 1982;<br>Beards 1967 |
| **Widder Formation (Hamilton Group)** | Area between Lake Erie and Lake Huron | Givetian | Shallow marine | 21 m (max.) | Predominantly soft, grey, calcareous shales; interbeds of argillaceous, crinoidal and nodular limestone; richly fossiliferous with brachiopods, bivalves, cephalopods, trilobites.<br>Lower contact is conformable with Hungry Hollow Fm. | Uyeno et al. 1982;<br>Winder and Sanford 1972 |

Table 20.8. Middle Devonian (Givetian) to Early Mississippian (Tournaisian) units in Ontario—Depositional Sequence 8 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Hungry Hollow Formation (Hamilton Group)** | Area between Lake Erie and Lake Huron | Givetian | Shallow marine | 2.0 m (max.) | Grey, soft shales. Brown argillaceous, coarse-grained, crinoidal and bioclastic limestone; trace fossils and corals.<br>Lower contact is sharp and disconformable(?) with Arkona Fm. | Uyeno et al. 1982;<br>Mitchell 1967;<br>Landing and Brett 1987 |
| **Arkona Formation (Hamilton Group)** | Area between Lake Erie and Lake Huron | Givetian | Shallow marine | 32 m (avg.) | Blue-grey, soft shales with occasional thin argillaceous limestone interbeds; relatively unfossiliferous except for local brachiopod concentrations.<br>Lower contact is sharp with Rockport Quarry Fm. | Uyeno et al. 1982 |
| **Rockport Quarry Formation (Hamilton Group)** | Area between Lake Erie and Lake Huron | Givetian | Shallow marine | 5.7 m (avg.) | Grey to grey-brown, fine-grained to micritic limestone with shale interbeds; shale content increases to southeast; grades laterally into Arkona Fm.<br>Lower contact is sharp with Bell Fm. | Sanford 1968;<br>Uyeno et al. 1982 |
| **Bell Formation (Hamilton Group)** | Area between Lake Erie and Lake Huron | Givetian | Shallow marine | 14.5 m (average) | Blue-grey, soft shales with thin limestone and organic-rich shale interbeds in lower part of unit; homogeneous and rich in shell (brachiopod) debris.<br>Lower contact is unconformity with Dundee or Marcellus Fms. (Sequence 7). | Uyeno et al. 1982;<br>Johnson et al. 1989;<br>Johnson et al. 1985 |
| **Northern Ontario** | | | | | | |
| **Long Rapids Formation** | Moose River and Hudson Bay basins | Frasnian to Famennian | Shallow marine environment with stratified water column; very shallow to marginal marine in Hudson Bay Basin | 30 m (avg.);<br>80 m (max.) | Black, siliciclastic, organic-rich shale; green-grey mudstone and minor limestone.<br>Tasmanites (small algal cysts); conodonts; burrows; fossilized plants.<br>Upper mb.: poorly consolidated, grey-green shale and minor mudstone interbeds.<br>Middle mb.: black, fissile, organic-rich shale with lesser organic-poor carbonate and mudstone beds.<br>Lower mb.: green-grey mudstone alternating with organic-rich shales and minor carbonate beds; microcrystalline calcareous concretions are abundant.<br>Lower contact is disconformble with Williams Island Fm. | Dyer and Crozier 1933;<br>Sanford and Norris 1975;<br>Telford 1984;<br>Bezys 1987;<br>Bezys and Risk 1990 |
| **Williams Island Formation** | Moose River and Hudson Bay basins | Givetian | Subtidal to supratidal environment | Total:<br>70 to 90 m<br>Upper mb.:<br>33 to 45 m<br>Lower mb.:<br>36 to 47 m | Upper mb.: brown to grey, argillaceous, oolitic and brecciated limestone with lesser dolostone.<br>Lower mb.: blue-grey shale with sandstone, gypsiferous shale, gysiferous sandstone and shale; brown, brecciated limestone interbeds.<br>Local basal unit of red, calcareous shale.<br>Lower contact is unconformity with Murray Island Fm. (Sequence 7). | Sanford and Norris 1975;<br>Telford 1988 |

with crinoidal debris, bryozoan, coral and bivalve material (Uyeno et al. 1982). The Widder–Ipperwash contact is not exposed, although subsurface records suggest it is quite sharp (Beards 1967). The Ipperwash Formation is poorly exposed in Ontario, with the only outcrop being a 1 m thick section at Ipperwash Beach (the type section), near Stony Point on Lake Huron (Uyeno et al. 1982). One unusual feature of the Ipperwash Beach outcrop is the occurrence of large burrow-like structures, up to 15 cm in diameter and several metres in length (Uyeno et al. 1982). The Ipperwash Formation is quite variable in thickness (2 to 13 m), possibly due to pronounced erosion during the hiatus preceding deposition of the overlying Kettle Point Formation (Uyeno et al. 1982). The upper contact of the Ipperwash Formation with the Kettle Point Formation is sharp and disconformable.

Overlying the unconformities which form the upper boundaries of the Hamilton Group in southern Ontario and the Williams Island Formation in the Hudson Platform, are organic-rich shales of the Kettle Point and Long Rapids formations, respectively. These deposits (including the Marcellus Formation of Sequence 7) are part of a Middle Devonian to Lower Mississippian black shale succession, which extends from Alabama and Oklahoma in the south, to Ontario and Michigan in the north, and as far west as Iowa (*see* Figures 20.40 and 20.41; Ettensohn and Barron 1981; Matthews 1983).

Although Middle Devonian black shale deposits such as the Marcellus Formation (*see* Sequence 7) were mainly restricted to the Appalachian Basin, black shale deposition during the Late Devonian occurred in most intracratonic basins of eastern North America. In Ontario, this deposition is represented by the Kettle Point Formation, which was deposited between the Michigan and Appalachian basins, and the Long Rapids Formation in the Moose River and Hudson Bay basins (Chart D, Map 2582, map case; *see* Figures 20.40, 20.41 and 20.42). Possible Early Mississippian black shale deposits in Ontario are represented by the Sunbury Formation of the Port Lambton Group.

The Kettle Point Formation represents a "compressed" interval of rocks as compared to equivalent strata of the Michigan and Appalachian basins proper. Biostratigraphic and lithostratigraphic evidence indicates that the Kettle Point Formation is approximately equivalent to the Antrim Shale (Michigan Basin) and the Ohio Shale (Appalachian Basin), but was deposited at a slower rate (Russell 1985). Figure 20.41 illustrates the stratigraphic relationship of the Kettle Point Formation with the approximately equivalent black shale units of eastern North America. Kettle Point shale deposition may have been more widespread over the Algonquin Arch, however, it is now only preserved in the Chatham Sag (*see* Figure 20.2; Uyeno et al. 1982; Russell 1985).

The Kettle Point Formation consists of siliciclastic, organic-rich shale, with minor organic-poor shale interbeds (Uyeno et al. 1982; Russell 1985). The formation ranges in thickness from 30 m in the Chatham Sag area, where it subcrops directly beneath Quaternary cover, to approximately 75 m near Sarnia, where it is disconformably overlain by the Port Lambton Group (Sanford 1968; Telford 1984). The Kettle Point Formation is markedly thinner than both the Ohio (over 300 m: Broadhead et al. 1982) and Antrim shales (about 200 m, Sanford 1968).

Large (up to 1.2 m) calcite concretions (Figure 20.43) are found in the lower one-third of the formation, well exposed at the only known outcrop of the formation, at Kettle Point on Lake Huron. The name "Kettle Point" originates from the similarity in shape of these concretions with cooking kettles. The early diagenetic origin of these concretions was recently documented by Coniglio and Cameron (1990).

Organic-rich intervals in the Kettle Point Formation tend to be laminated and contain 3 to 15% total organic carbon, whereas the organic-poor interbeds are commonly bioturbated and contain less than 2% total organic carbon (Russell and Barker 1983; Armstrong 1986). Organic-poor interbeds are concentrated in a basal 10 to 20 m interval and in a thin interval in the upper half of the formation (Russell and Barker 1983). These beds predominate towards the

**Figure 20.43.** Large (approximately 1 m diameter) carbonate concretion in outcrop of black (and minor green) shales of the Kettle Point Formation, at Kettle Point, on the shore of Lake Huron (photograph *courtesy of* D. Armstrong).

Algonquin Arch, suggesting it was a positive feature separating the Michigan and Appalachian basins during the Late Devonian (Russell 1985).

Although generally barren of macrofossils, except for rare lingulid brachiopods and fish remains, the small algal form or cyst *Tasmanites* is abundant and ubiquitous throughout the Kettle Point Formation. Other microfauna, such as pyritized radiolaria and sponge spicules and conodonts, have also been identified (Winder 1966; Armstrong 1986). Plant remains are preserved as either carbonized or silicified woody material (Armstrong 1986). The stratigraphically restricted algal form *Foerstia*, identified in this unit by Winder (1966), was used for regional correlation by Russell (1985). The depositional environment interpreted for the Kettle Point Formation is presented in the discussion of the Long Rapids Formation of northern Ontario.

Towards the end of the Devonian, sedimentation in Ontario took place under more normal marine conditions with the deposition of the Port Lambton Group (Winder and Sanford 1972). This group of clastic rock units consists mainly of grey and black shales and sandstones (Winder and Sanford 1972). In Ontario, these units are restricted to the subsurface of Moore and Sombra townships of Lambton County, bordering the St. Clair River (Sanford 1968; Uyeno et al. 1982). The contact between the Port Lambton Group and the underlying Kettle Point Formation is sharp (Beards 1967; Johnson et al. 1985) and locally erosional, where sandstone channels (of the Berea Formation) scour into the Kettle Point Formation (Sanford 1968).

The precise age of the Port Lambton strata in Ontario is equivocal (Uyeno et al. 1982). Biostratigraphic evidence indicates either a Late Devonian or Early Mississippian age (Uyeno et al. 1982), and thus, the systemic boundary may occur within the group.

Three formations are defined within the Port Lambton Group in the Michigan Basin, where their thicknesses warrant formational status (*see* Figure 20.42 and Table 20.8; Chart D, Map 2582, map case). Approximately 30 m of unfossiliferous shales have been assigned to the Bedford Formation in Lambton County (Sanford 1968; Johnson et al. 1985). The Bedford Formation is overlain gradationally to sharply by the interbedded sandstones, siltstones and shales of the Berea Formation. Anomalous thicknesses of the Berea Formation, up to 60 m, are associated with deposition of "shoe-string" sandstones, which locally cut down through the underlying Bedford Formation and into the Kettle Point Formation (Sanford 1968). The Sunbury Formation, consisting of up to 20 m of black shales in Ontario, sharply and disconformably overlies the Berea Formation (Uyeno et al. 1982). The Sunbury Formation is unconformably overlain by Quaternary sediments in southwestern Ontario (Johnson et al. 1985); no younger Paleozoic strata have been preserved in the province.

## NORTHERN ONTARIO

On the Hudson Platform, the Williams Island Formation is stratigraphically equivalent to the Hamilton Group (Chart D, Map 2582, map case; *see* Figure 20.42) and disconformably overlies the Murray Island Formation of Sequence 7. The lithologic and faunal similarities of the Williams Island Formation to the Hamilton Group suggest a physical connection (i.e., seaway) between the 2 regions at this time (Sanford and Norris 1975). Outcrops of the Williams Island Formation are restricted to riverbank exposures in the eastern part of the Moose River Basin (Figure 20.44; Sanford and Norris 1975; Telford 1988).

The Williams Island Formation has been subdivided into 2 members (Sanford and Norris 1975). The lower member consists of shales, with lesser sandstones and limestones, and minor evaporites (*see* Table 20.8; Telford 1988). The upper member consists of a mixture of argillaceous, oolitic, dolomitic and brecciated limestones with lesser calcareous shales and dolostones (*see* Figure 20.44).

The clastic beds in the lower member may have been derived from the uplift and erosion of cratonic highlands (i.e., arches) around the Hudson Platform (Sanford and Norris 1975). The presence of evaporites indicates deposition in a supratidal environment; however, the localized presence of normal marine fossils in these sediments indicates intermittent subtidal deposition (Sanford and Norris 1975; Telford 1988).

The Long Rapids Formation is the youngest Paleozoic formation in the Hudson Platform (Chart D, Map 2582, map case; Figure 20.42). It disconformably overlies the William Island Formation and underlies Mesozoic and Quaternary sediments. The unit has only limited exposure, along the Abitibi (Figure 20.45) and Mattagami rivers, and is known primarily from subsurface records (Sanford and Norris 1975). Recent subsurface studies (Telford 1984) indicate a more limited present day distribution for the Long Rapids Formation than previously thought. In the Moose River Basin, the unit reaches a maximum thickness of about 80 m (Onakawana B borehole, Sanderson and Telford 1985; Bezys 1987, 1989).

In the Moose River Basin, the Long Rapids Formation is lithologically similar to the Kettle Point Formation, consisting of black, organic-rich shale with lesser, organic-poor carbonate and mudstone interbeds (Bezys and Risk 1990).

Figure 20.44. Thin- to medium-bedded limestone of the upper member of the Williams Island Formation, exposed on Williams Island, on the Abitibi River (photograph courtesy of P. Telford).

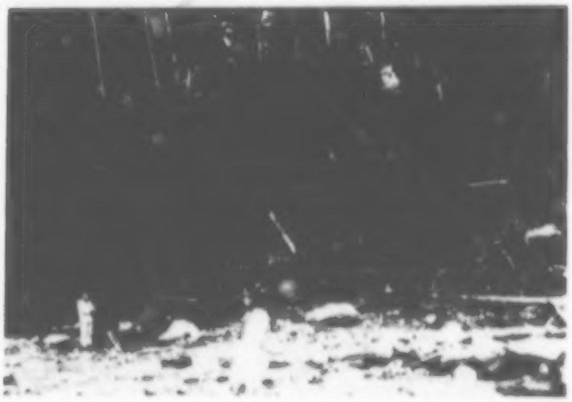

**Figure 20.45.** Black shales of the middle member of the Long Rapids Formation in an outcrop on the Abitibi River (photograph *courtesy of* R. Bezys).

Three informal members have been identified (*see* Table 20.8), the most organic-rich being the shales of the middle member (*see* Figure 20.45). Microcrystalline, calcareous concretions are frequently found within the lower part of the formation (Bezys and Risk 1990). Generally, these concretions are smaller than those of the Kettle Point Formation and do not display the same radiating crystal growth.

The algal form or cyst, *Tasmanites*, has been identified in the Long Rapids Formation (Bezys and Risk 1990), as have conodonts (observed on bedding plane surfaces), and, in the concretions, burrows, rare fossils and fossilized plant material (Bezys and Risk 1990). Trace fossils occurring in the mudstone beds of this unit constrain water depth and oxygen content in interpretations of the depositional environment (Bezys 1987).

Both the Long Rapids and Kettle Point formations are interpreted to have been deposited in a relatively shallow marine setting with a stratified (i.e., density, temperature and oxygen content) water column (Russell 1985; Telford 1988; Bezys and Risk 1990). In this depositional model, organic matter in the surface photic zone would be protected from aerobic decay as it settled through the anoxic (or anaerobic) water column. Vertical movement of the pycnocline (the thin zone which separates oxygenated surface waters from denser, anoxic bottom waters), perhaps due to fluctuation in surface biological productivity, would cause the pycnocline to impinge on the seafloor periodically, promoting bioturbation and the deposition of organic-poor, green-grey shales or carbonates. Recent studies of the Black Sea (Calvert 1990), long considered to be the modern analog for organic-rich sedimentation in a stratified sea (Demaison and Moore 1980), indicate that surface biological productivity may be a more significant factor than stratification of the water column. This suggests that the currently accepted models for Devonian black shale deposition need to be re-evaluated.

Strata intersected in boreholes in the Hudson Bay Basin and tentatively assigned to the Long Rapids Formation consist of evaporitic mudstones and reddish siltstones and sandstones (Norris 1986; Thorpe 1988). These deposits suggest that very shallow water conditions existed in the Hudson Bay Basin during the Late Devonian. They give some support to the interpretation of relatively shallow water deposition for the Long Rapids Formation in the adjacent Moose River Basin (Telford 1988).

## Economic Geology

Significant resources of Sequence 8 consist of shales of the Hamilton Group and hydrocarbon resources of the Long Rapids and the Kettle Point formations.

Of the Hamilton Group units, the Arkona Formation has the best potential as a source of shale for ceramics, brick and tile manufacturing (Guillet and Joyce 1987). Shale has been extracted from pits along the Ausable River near Thedford (Guillet and Joyce 1987). Minor conventional hydrocarbons are also present in the Hamilton Group (*see* Economic Geology section of Sequence 7).

During the mid 1980s, the OGS investigated the oil shale resources of the Marcellus, Kettle Point and Long Rapids formations (Barker et al. 1983; Johnson et al. 1983a, 1983b, 1985, 1989; Johnson 1983, 1985; Telford 1984). Results of the Marcellus investigation are noted in the Economic Geology section of Sequence 7. The Long Rapids Formation contains similar oil yields to the Kettle Point Formation, although the remoteness of the former greatly decreases its development potential. (For more information on the Long Rapids Formation *see* Bezys and Risk 1990.)

Kettle Point Formation samples were analyzed for total organic carbon content, oil yield on pyrolysis, Fischer Assay and other organic geochemical parameters (Barker et al. 1983; Barker 1985 ). Using conventional retorting methods (the Fischer Assay method), the formation produced up to 60 litres per tonne (13 gallons per ton). Retorting in a hydrogen-rich environment (i.e., the HYTORT process) increased yields to 130 litres per tonne (Barker 1985). The most organic-rich zones are located in the upper and lower parts of the Kettle Point Formation (Russell 1985) and are readily delineated using boreholes and geophysical methods (Telford 1984; Russell 1985). Proximity of the Kettle Point shales to the petrochemical facilities at Sarnia enhances their potential resource value.

The black shale units of eastern North America have collectively been termed the Eastern Gas Shales, because, where they are more deeply buried, such as in the Appalachian Basin, they are significant producers of natural gas (e.g., Janka and Dennison 1980; Broadhead et al. 1982). The Eastern Gas Shales have also been assessed as a major shale oil resource in the United States. A survey by the Institute of Gas Technology, reported by Janka and Dennison (1980), estimated a surface-mineable shale oil resource in the eastern United States of 423 billion barrels of oil, based on processing the shales using the patented HYTORT method.

The Kettle Point Formation, like most other organic-rich shales (Vine and Tourtelot 1970), is also enriched in a number of minor and trace metals (Armstrong 1986). These elements, which include sulphur, uranium, nickel, vanadium, molybdenum, copper, lead, zinc and arsenic, may enhance the economic potential of the Kettle Point shales, if they can be readily extracted in the shale oil process. They may, however, also pose an environmental hazard if not appropriately processed.

## DEPOSITIONAL SEQUENCE 9: MISSISSIPPIAN (TOURNAISIAN) TO PERMIAN (ASSELIAN) (?)

### Overview

Although there are no proven Sequence 9 strata preserved in Ontario, there is some evidence that sediments were deposited, in at least southern Ontario, during this time. Sequence 9 strata in the adjacent areas of the Michigan and Appalachian Basins are dominated by clastic sediments derived from the Alleghanian Orogeny, with lesser amounts of limestone and coal. The lower boundary of this sequence is marked by the return to orogen-derived clastic sedimentation brought on by early Alleghanian or late Acadian tectonism. The upper boundary is marked by a widespread erosional unconformity brought about by regional-scale uplift and erosion throughout the eastern North American continent.

Sediments of Sequence 9 were probably more widespread, but removed by erosion during the Mesozoic. The marine connection between the St. Lawrence and Hudson platforms was permanently severed by uplift related to either latest Acadian or earliest Alleghanian tectonism which occurred at the close of the Devonian. The Alleghanian Orogeny, which prevailed throughout much of the Mississippian(?), Pennsylvanian and into the Permian(?) (Chart D, Map 2582, map case), yielded thick deposits of marine and nonmarine clastic rocks, limestones and coal in the Appalachian and Michigan basins (Cohee et al. 1951; de Witt 1975). These strata were likely continuous across the Algonquin and Findlay arches in Ontario, but were subsequently removed from these areas by uplift and erosion (Figure 10, Map 2582, map case).

The clastic wedges in the Appalachian and Michigan basins were largely orogen derived, with some detritus in the latter region likely derived from cratonic sources, presumably the adjacent Fraserdale and Frontenac arches. Lesser, but extensive, carbonate sediments were deposited by a marine transgression during the middle Mississippian (Milici and de Witt 1988). The 2 pulses of deposition during the Carboniferous were separated by an episode of cratonic uplift and erosion (Fisher et al. 1988; Chart D, Map 2582, map case).

This sequence was terminated by large-scale uplift and erosion which lasted until the middle Mesozoic (*see* Sequence 10) and resulted in a large gap in the sedimentary record in Ontario (Chart D, Map 2582, map case).

### Geology

Within Ontario, there are virtually no strata representing the Mississippian through to Triassic periods. The Port Lambton Group of southwestern Ontario (*see* Sequence 8) may be in part Early Mississippian (Tournaisian) in age (Fisher et al. 1988); although this is still contentious (Uyeno et al. 1982). In the central Hudson Bay Basin, boreholes have intersected marine clastic beds and evaporites of possible Pennsylvanian age (Thorpe 1988; not shown on Chart D, Map 2582, map case); no equivalents are known in the Moose River Basin. Evidence for the previous existence of Sequence 9 age strata in southern Ontario is based on data from a number of fields of investigation, as presented below.

Adjacent to Ontario, considerable thicknesses of Sequence 9 deposits are known: over 700 m in the Michigan Basin (Fisher et al. 1988) and over 2400 m in the northern Appalachian Basin (Milici and de Witt 1988); however, in the western part of the Appalachian Basin, a thickness of 150 to 300 m is more typical. Although there are no Carboniferous strata preserved in Ontario connecting these 2 major depocentres, lithostratigraphic correlation suggests that sedimentation was continuous between the 2 basins (Figure 10 and Chart D, Map 2582, map case). Comparison of Carboniferous stratigraphy preserved in the Michigan and Appalachian basins (Table 20.9; Fisher et al. 1988; Milici and de Witt 1988) shows that 2 approximately simultaneous pulses of mainly clastic sedimentation occurred in both basins during the Mississippian and the Pennsylvanian (and into Permian?).

In the central part of the Michigan Basin, about 520 m of Mississippian strata, consisting of marine shales and sandstones capped by a limestone unit, is present (*see* Table 20.9; Fisher et al. 1988). The highly variable thickness of the uppermost unit (Bayport Limestone) of this interval is interpreted as being due to Late Mississippian erosion. The overlying Pennsylvanian strata in Michigan consist of sandstone, and interbedded shales, thin coals and limestones (*see* Table 20.9; Fisher et al. 1988).

Two thicker intervals of similar sediments are preserved in the Carboniferous stratigraphy of the Appalachian Basin (*see* Table 20.9; Milici and de Witt 1988) and presently extend northwestward, almost to the southern shore of Lake Erie. During the Early Mississippian, a dominantly siliciclastic succession was deposited in a prograding deltaic environment (Milici and de Witt 1988). A major transgression in the Late Mississippian deposited extensive carbonate sediments (Maxville Limestone). Following widespread uplift and erosion, a second pulse of orogen-derived clastic sediments were shed from the Alleghanian Orogeny during the Pennsylvanian. These predominantly clastic sediments included coal and limestone interbeds and were deposited in a complex interplay of terrestrial-deltaic to nearshore marine environments (Milici and de Witt 1988).

The prior existence of Carboniferous (or younger?) deposits in Ontario can also be inferred, and their approximate original thickness calculated, by determining the burial history of preserved strata. Thermal maturation indicators, that is, geological materials that change properties (e.g., colour) when exposed, at depth, over a period of time, to temperatures above their original depositional temperature, are commonly used for such purposes

**Table 20.9.** Lithostratigraphic comparison of Mississippian to Permian (Tournaisian to Asselian) units of Michigan and Ohio—Depositional Sequence 9 (described in descending stratigraphic order; *modified from* Fisher et al. 1988; Milici and de Witt 1988).

| Unit | Age/Stage | Rock Type(s) |
|---|---|---|
| **Michigan** | | |
| Grand River Formation | Westphalian | Sandstone |
| Saginaw Formation | Namurian to Westphalian | Sandstone, shale, coal, limestone |
| Bayport Limestone | Visean | Limestone |
| Michigan Formation | Visean | Shale and sandstone |
| Marshall Formation | Visean | Shale and sandstone |
| Coldwater Formation | Tournaisian to Visean | Shale with minor limestone and siltstone |
| **Ohio** | | |
| Dunkard Group | Stephanian to Asselian (?) | Sandstone, conglomerate, mudstone |
| Monongahela Group | Stephanian | Shale and limestone; coal, mudstone, sandstone |
| Conemaugh Group | Stephanian | Sandstone, shale mudstone, coal, limestone |
| Allegheny Group | Westphalian | Sandstone, siltstone, shale, coal, limestone |
| Pottsville Group | Namurian to Westphalian | Conglomerate, sandstone shale, coal, limestone |
| Sharon Conglomerate | Namurian | Conglomerate |
| Maxville Limestone | Visean to Namurian | Limestone with minor shale and dolostone |
| Logan Sandstone | Visean | Sandstone |
| Cuyahoga Group | Tournaisian to Visean | Sandstone, siltstone mudstone |

(e.g., Héroux et al. 1979). These indicators include hydrocarbon maturation levels, conodont alteration indices, acritarch colouration indicators, vitrinite reflection levels, oxygen isotopic ratios and fluid inclusion homogenization temperatures (Epstein et al. 1977; Héroux et al. 1979). Experience with these indicators has permitted assignment of burial depths based on temperature ranges obtained from the indicators (e.g., Legall et al. 1981; Friedman 1987).

Legall et al. (1981) determined from conodont and acritarch colour alteration indices that the Middle Devonian strata of southwestern Ontario were covered by 1300 to 2000 m of younger sediments, all of which, except for less than 100 m of Upper Devonian (*see* Sequence 8) strata, have been removed by (post–Devonian) erosion. Fisher et al. (1988) reported vitrinite reflectances of Pennsylvanian age coals in Michigan that indicate a minimum of 900 m of overlying (probably Pennsylvanian) sediments, which have been subsequently eroded. Combining these estimates with the existing preserved Carboniferous interval in the Michigan Basin (700 m; Fisher et al. 1988) yields a probable thickness similar to that proposed by Legall et al. (1981) for southwestern Ontario.

In the northwestern Appalachian Basin, in western New York, fluid inclusion homogenization temperatures and conodont alteration indices indicate a depth of burial for Middle Silurian strata of approximately 5 km (Friedman 1987). Subtracting the 1300 m average thickness for overlying Silurian and Devonian strata (Winder and Sanford 1972) results in an estimated thickness of overburden of 3700 m for the northwestern Appalachian Basin. The thicker Carboniferous interval in the Appalachian Basin (relative to the above estimates for southwestern Ontario) is not unexpected considering this basin's proximity to the Alleghanian Orogen.

Thermal maturation indicators suggest higher burial temperatures for strata in eastern Ontario (east of the Frontenac Arch), increasing dramatically to about 185° to 300°C in the Montreal area (Legall et al. 1981). In addition, analysis of oil shales of the Eastview Member (Lindsay Formation; *see* Sequence 2) in eastern Ontario show much higher maturation levels than are present in equivalent oil shales (Collingwood Member of Lindsay Formation) in south-central Ontario (Barker et al. 1983). These higher maturation levels likely do not reflect greater burial depths, but have instead been attributed to more intensive local faulting and intrusive igneous activity, possibly related to a high temperature mantle plume, or hot spot, occurring in this area during the Cretaceous to Early Tertiary (Crough 1981). This plume is believed to have caused doming of the craton, thereby encouraging erosion of the Paleozoic cover.

Although the youngest strata now present in eastern Ontario belong to the Queenston Formation (*see* Figure 20.8), Devonian fragments in the Mesozoic Sainte-Helene Island diatreme breccia at Montreal indicate that rocks of at least Devonian age may have been present in eastern Ontario (*see* Sequences 7 and 8; Boucot et al. 1986). Fragments of younger strata may also be present in this breccia, but these may not have been identified because of their sparse fossil content.

The thermal history of Sequence 8 and older strata indicates substantial depths of burial. However, evidence that the overlying, and subsequently eroded, material was Mississippian and Pennsylvanian in age is based solely on the presence of strata of that age in the adjoining Appalachian and Michigan basins.

# DEPOSITIONAL SEQUENCE 10: MIDDLE JURASSIC (BAJOCIAN) TO LATE CRETACEOUS (CENOMANIAN)

## Overview

The stabilization of the Appalachian Orogen in post-Carboniferous or Permian time was followed by the onset of a completely new tectonic cycle, which began with rifting in eastern North America during the Late Triassic, and was followed by, in turn, continental separation and the eventual formation of the present Atlantic Ocean during the Jurassic to mid–Cretaceous (Hatcher 1989; Sanford, in press). The strata of Sequence 10, consisting of widely scattered nonmarine to marine, craton-derived clastic sediments, overlie an extensive unconformity associated with widespread uplift at the close of the Alleghanian Orogeny. Deposition of Sequence 10 sediments was followed by a major erosional episode which has continued to the present.

The Mesozoic rifting process was similar in many respects to the earlier rifting and continental separation during the Hadrynian to Early Cambrian (*see* Sequence 1), except that its greatest effects were registered farther south, close to the present continental margin. However, the Canadian craton was not entirely unaffected by the activities occurring on the eastern margin of North America. There is substantial evidence on the craton of Precambrian basement rejuvenation in the form of uplift and block faulting (Sanford et al. 1985). Additionally, the emplacement of dikes and plutons of Middle Jurassic to Early Cretaceous age over wide regions of the Canadian Shield (*see* Easton, this volume; Sage, this volume) reflect the extent of magmatism associated with the rifting (Manspeizer et al. 1989).

The presence of Upper Jurassic (Kimmeridgian) age red bed clastic sediments of nonmarine origin in the Michigan Basin (Sanford et al. 1979; Fisher et al. 1988) and Middle Jurassic (Bajocian) age clastic sediments (the Mistuskwia Beds) in the Moose River Basin (Telford and Long 1986) points to reactivation and uplift of the Fraserdale Arch during these times (Figure 11, Map 2582, map case). Emplacement of ultramafic dikes along the Frontenac Arch (e.g., Barnett et al. 1984) suggest that this arch was also tectonically active during this time.

Mesozoic rocks have a relatively limited distribution on the Hudson Platform. Marine clastic sediments of Cretaceous (Aptian to Cenomanian) age are widely scattered, and occur in the Hudson Bay Basin and to the north beneath Hudson Strait and Foxe Channel (Sanford and Grant 1990). These recently identified beds (Evans Strait Formation) are assumed to have been deposited in a marine seaway that extended through Hudson Strait and Hudson Bay, connecting the Labrador Shelf in the east with the Interior Platform in the west (Sanford and Grant 1990). Onshore, the distribution of these beds is confined to a small area of northern Ontario bordering Manitoba (Map 2541, map case).

In the Moose River Basin, Middle to Upper Albian fluvial and minor lacustrine deposits (Mattagami Formation; *see* Map 2543, map case) are remnants of a large, northwest-flowing river system that once drained large areas of the Canadian Shield (Try et al. 1984; Telford and Long 1986; Telford et al. 1991).

Following the deposition of these Mesozoic units in Ontario, all the bedrock of the province appears to have undergone a period of erosion which has continued unabated to the present. This erosion was particularly severe during the last 2 million years with the onset of Quaternary glaciation (*see* Barnett, this volume).

Lithological details and a stratigraphic chart of the units present in Sequence 10 are shown in Table 20.10 and Figure 20.46, respectively. The present distribution of these units is shown in Maps 2541 and 2543 (map case). Distribution of the sequence as a whole is shown on Maps 2575 and 2577 (map case). The estimated pre-erosion maximum distribution of this sequence is indicated on Figure 11, Map 2582 (map case).

## Geology

Mesozoic deposition in Ontario is represented by a thin veneer of sediment on the Hudson Platform and a minor remnant (not shown in Map 2542, map case) of the Interior Platform succession in the Rainy River–Fort Frances area (Zippi and Bajc 1990). Whether such sediments once blanketed the remainder of the province is not known. However, widespread Mesozoic sedimentation is indicated by the presence of Jurassic red bed clastic sediments in central Michigan (Fisher et al. 1988), and by the occurrence of Jurassic to Cretaceous age hydrocarbons in anthraxolite (a black coal-like substance) found in fractures and veins (Hayatsu et al. 1983) and Cretaceous spores in the iron ores at Atikokan (Machado 1987). The limited distribution of Mesozoic sediments today may be attributed to their relatively unconsolidated and friable nature, which made them susceptible to post-depositional erosion.

Overall, Mesozoic time in Ontario was a period of extensive erosion with only a minor preserved record of deposition. Three formations of Mesozoic age are present in the province (*see* Figure 20.46; Chart D, Map 2582, map case): the Middle Jurassic Mistuskwia Beds and the Lower Cretaceous Mattagami Formation and the Lower to Upper Cretaceous Evans Strait Formation. Of these 3 units, only the Mattagami Formation is exposed (although poorly) at surface, whereas the other 2 are known only from borehole records and offshore seismic profiles (Sanford and Grant 1990).

The Mistuskwia Beds (Telford 1982) of the central Moose River Basin consist of varicoloured calcareous clays with thinner beds of unconsolidated calcareous quartz sands (*see* Table 20.10). The unit is difficult to distinguish lithologically from the overlying Cretaceous or Pleistocene material. Palynological analysis, which originally determined the existence of these beds in OGS drill holes (*see* Telford et al. 1975, 1991; Norris 1977, 1982; Norris and Zippi 1986), remains the most reliable means of distinguishing this unit from underlying and overlying units (Telford and Long 1986). As this unit occurs only in the subsurface, it is not shown on the bedrock or tectonic maps (*see* Maps 2543 and 2577, map case); however, its location is shown on the sequence distribution map (Figure 11, Map 2582, map case). A more accurate representation of the distribution of the Mistuskwia Beds is found in Telford et al. (1991).

The base of the Mistuskwia Beds has been identified in only 1 drill core, OGS–75–02 (Telford et al. 1975). Here, the basal 2 m of the unit is a conglomerate (*see* Table 20.10; Telford 1982). The contact with the underlying limestones of the Williams Island Formation (of Sequence 8) is sharp and disconformable, representing a gap of approximately 190 million years (Norris 1982). The upper contact with the overlying Cretaceous or Pleistocene deposits appears to be sharp and disconformable.

The Mistuskwia Beds are interpreted as a lacustrine deposit (Telford and Long 1986). Clast types indicate local (i.e., cratonic) sources supplied the sediments (*see* Table 20.10; Hamblin 1982). The interval in drill hole OGS–75–03 shows an upward gradation from mud to mixed fine and coarse sand, and then finally, to medium to very fine sand and is interpreted as a prograding shoreline (lacustrine) environment. In drill hole OGS–75–02, the presence of thin laminae of green mudstone above the basal conglomerate suggest a nearshore, prodeltaic setting (Telford et al. 1991).

The Lower Cretaceous Mattagami Formation is more extensive than the underlying Mistuskwia Beds, covering approximately 7000 km² of the central and southern Moose River Basin (Maps 2543 and 2577; Figure 11, Map 2582, map case). The unit unconformably overlies Precambrian, Silurian, Devonian (Williams Island Formation) and Jurassic (Mistuskwia Beds) strata. The topographic irregularity of this contact, combined with pre–Pleistocene erosion of the upper surface of the Mattagami Formation, has resulted in highly variable thicknesses (14 to 166 m) for the unit (Try et al. 1984; Telford and Long 1986). The formation is clearly a remnant of a formerly more extensive cover (Figure 11, Map 2582, map case). Equivalent strata are known in Quebec and Manitoba (Remick et al. 1963; Telford and Long 1986) and Cretaceous strata occur in an outlier (not shown on Figure 11, Map 2582, map case) at Limestone Rapids, 56 km west of the main depositional area, in the Moose River Basin (Telford et al. 1991). Additionally, in Cargill Township (near Kapuskasing), silica sand and clay deposits preserved in karstic topography associated with the weathering of carbonatite complexes (Vos 1981; Telford and Long 1986) are similar to sediments of the Mattagami Formation and are believed to be Cretaceous in age (Norris 1982; Norris et al. 1980).

Outcrops of the Mattagami Formation are limited to small exposures along the Mattagami, Missinaibi and Abitibi rivers and associated tributaries. The most extensive exposures occur along Adam Creek, a diversion channel of the Ontario Hydro dam across the Mattagami River at Long Rapids.

**Table 20.10.** Middle Jurassic (Bajocian) to Late Cretaceous (Cenomanian) units in Ontario—Depositional Sequence 10 (described according to geographic location and in order of youngest to oldest unit).

| Unit | Distribution | Age/Stage | Depositional Environment | Thickness Range | Description | Selected References |
|---|---|---|---|---|---|---|
| **Northern Ontario** | | | | | | |
| Evans Strait Formation | Hudson Bay Basin | Aptian to Cenomanian | Shallow marine | 150 m (max. in Hudson Bay) | Interbedded sand and shales. | Grant and Sanford 1988; Sanford and Grant 1990 |
| Mattagami Formation | Central and southern Moose River Basin | Albian | Anastomosing river system complete with flood plain, lacustrine, swamp and bog components | 14 to 166 m | Weakly consolidated kaolinitic mudrock with lesser silica sand, gravel and lignite. Mudstones are grey, black, white, yellow or red and may be organic-rich and laminated, or may contain pebbles. Silica sand is white, massive, flat bedded or planar cross-stratified and associated with stacked gravel or silt deposits. Upper contact with Pleistocene deposits is difficult to determine.<br>Lower contact is sharp and disconformable with Mistuskwia Fm. and unconformable with older strata. | Try et al. 1984; Telford et al. 1991; Vos 1982a, 1982b; Telford and Verma 1982; Watts et al. 1983, 1984 |
| Mistuskwia Beds | Central Moose River Basin | Bojacian to Bathonian | Lacustrine deposits with prograding shoreline, nearshore, prodeltaic setting | 19 m (max.) (incomplete section) | Varicoloured (grey, brown, green, pink, red), calcareous clays with thin beds of medium-grained, unconsolidated calcareous quartz sands. Basal 2 m is a conglomerate with abundant limestone and red sandstone fragments, pyrite concretions, quartz, chert and volcanic pebbles in sandy or silty matrix.<br>Upper contact is sharp and disconformable with either Mattagami Fm. or Pleistocene deposits.<br>Lower contact is sharp and disconformable with Williams Island Fm. (Sequence 8) | Telford 1982; Telford et al. 1975; Telford and Long 1986; Hamblin 1982; Telford et al. 1991 |

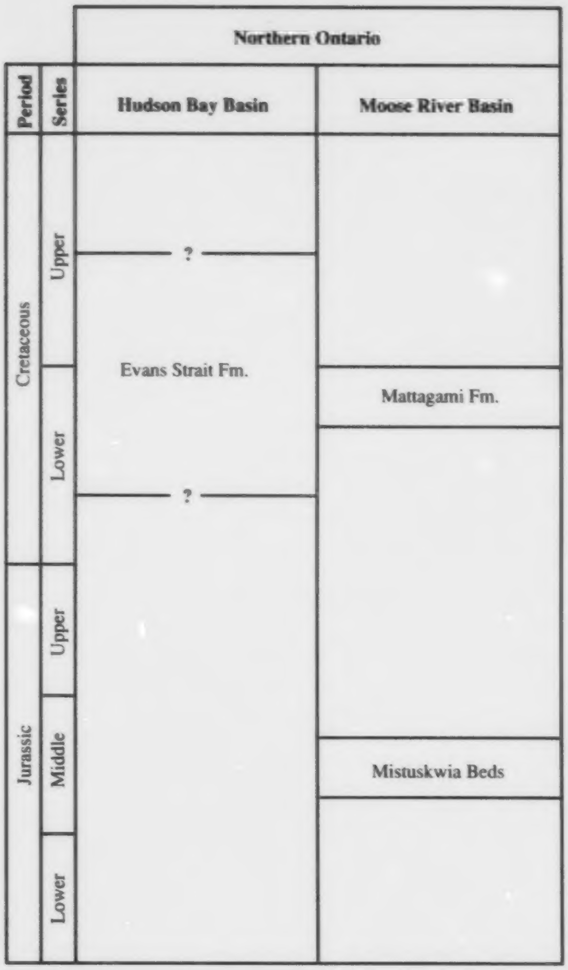

**Figure 20.46.** Stratigraphic units of Sequence 10 (? indicates uncertain age).

Interest in lignite (Figure 20.47) and other industrial minerals present in the Mattagami Formation (*see* Economic Geology section) led to a number of shallow drilling programs which were carried out between 1974 and 1988 (Telford and Verma 1982; Vos 1982a, 1982b; Watts, Griffis and McOuat 1983, 1984; Telford et al. 1991). Therefore, the subsurface distribution and character of this unit is fairly well understood. The Mattagami Formation primarily consists of weakly consolidated kaolinitic mudrock (clays) with lesser amounts of silica sand, gravel and lignite. The silica sands are mainly associated with stacked sand and gravel deposits, forming units up to 20 m thick (Try et al. 1984). Lithified, large pebble conglomerates are commonly associated with these sand and gravel deposits (Try et al. 1984; Telford et al. 1991).

The lower contact of the Mattagami Formation with the Mistuskwia Beds is sharp and disconformable, yet sometimes difficult to determine without palynological analysis. Similarly, the upper contact with Pleistocene and Recent material is, in some localities, difficult to determine due to the unconsolidated nature of the Mattagami Formation and its incorporation into the overlying deposits.

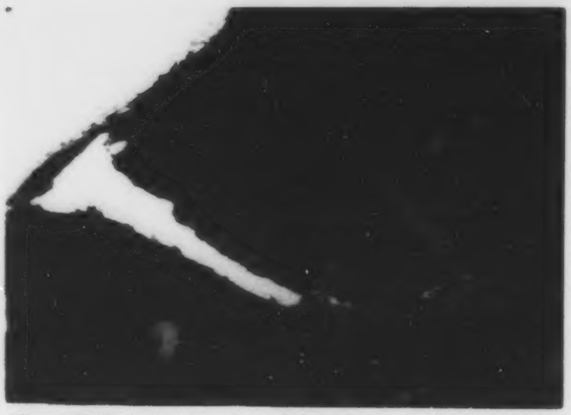

**Figure 20.47.** Aerial view of an old test pit at the Onakawana lignite field, James Bay Lowland. A stock pile of lignite excavated from this pit can be seen adjacent to the Abitibi River, near the centre of photograph (photograph *courtesy of* P. Telford).

The lithological characteristics of the Mattagami Formation suggest that it was deposited in an anastomosing segment of a large, northwestwardly flowing river system, complete with flood plain, lacustrine, swamp and bog components (Try et al. 1984; Telford et al. 1991). Precambrian highlands in the south were the main source of sediment for this fluvial system.

The Evans Strait Formation, consisting of sands and shales, has only recently been identified in the Hudson Bay Basin in 2 boreholes and in offshore seismic surveys (Grant and Sanford 1988; Sanford and Grant 1990). This unit appears to be preserved in depressions on the seafloor, in a concentric pattern, geographically restricted (Figure 11, Map 2582, map case) to subcrop areas of the Middle Devonian Moose River Formation (*see* Sequence 7). These depressions likely resulted from dissolution of evaporites in the Moose River Formation, which protected the Evans Strait Formation from erosion. Offshore seismic profiles indicate that this unit may be up to 150 m thick under Hudson Bay (Sanford and Grant 1990).

The Evans Strait Formation strata in Ontario have been assigned to a clastic marine succession of the same name that has been identified in Evans Strait and in the Southampton and Ungava basins to the northeast of Hudson Bay (Sanford and Grant 1990). The great thickness of the unit in these latter areas (up to 1000 m thick), combined with the protected occurrence of the Evans Strait Formation in Ontario, suggests that its original regional distribution has been significantly reduced by post-depositional (latest Cretaceous?) erosion.

Until recently, the present day distribution of Mesozoic strata in Ontario was believed to be restricted to the Hudson Platform. However, during 1989, drilling by the OGS in the Rainy River–Fort Frances area encountered a 20 m thick lens of Mesozoic sediment below 30 m of Quaternary cover (Zippi and Bajc 1990). Palynological studies indicate that this sediment is of Aptian to Cenomanian age (Early to Late

Cretaceous). Too small to show on the accompanying maps (Maps 2542 and 2576, map case), this deposit was undoubtedly associated with the more extensive Cretaceous sediments of the Interior Platform to the west.

### MESOZOIC IGNEOUS ACTIVITY

The crustal extension experienced along the eastern margin of the North American craton during the Jurassic and Cretaceous was accompanied by extensive magmatism, resulting in the widespread emplacement of dikes and other intrusions (Manspeizer et al. 1989). In southern Ontario, Middle Jurassic dikes, dated at 170 to 173 million years by Barnett et al. (1984) and 176±3 million years by Heaman (1989), cut Ordovician strata near Picton and Varty Lake in the Belleville–Kingston area (Map 2544; Chart D, Map 2582, map case). These dikes are spacially associated with faults (i.e., Picton Fault) and are believed to be related to the reactivation of fault systems during the rifting events on the eastern continental margin during the Jurassic (Barnett et al. 1984; McFall 1990).

In the southern part of Moose River Basin, Devonian strata are cut by Middle Jurassic to Early Cretaceous(?) intrusions dated at 150 to 180 million years (*see* Sage, this volume); a date of 128±18 million years was reported by Sanford and Norris (1975). A representative sampling of these intrusions is shown on Map 2543 (map case) and their age ranges are plotted on Chart D, Map 2582 (map case). These mafic to ultramafic intrusions are of lamprophyric and kimberlitic composition and have been an exploration target for diamonds (*see* Sage, this volume; Brummer 1978).

## Economic Geology

The Mattagami Formation contains significant deposits of lignite (low-grade coal), silica sand and kaolinitic clays (Vos 1982a, 1982b; Telford and Long 1986; Telford et al. 1991); however, the remoteness of their location has limited the development of these deposits. The resource potential of the other Mesozoic units in Ontario is unknown.

The silica sand and kaolinitic clay deposits occur in an interbedded succession and are seen in outcrop along the Missinaibi and Mattagami rivers (Guillet and Joyce 1987). At some locations, the silica sand is of high chemical purity and has remained a target of exploration interests (Vos 1982a, 1982b). Locally, the kaolinitic clays are of sufficient purity to be suitable for unrefined fire and china clay (Vos 1982a, 1982b).

Lignite was first reported as a mineral resource in the Moose River Basin in the 19th century (e.g., Bell 1877), although it was well known, from outcrops in the Onakawana area, to early trappers and the Hudson Bay Trading post at Moose Factory as early as the 1650s (Telford and Verma 1982). Lignite deposits are located at Onakawana (on the Abitibi River; *see* Figure 20.47) in the eastern part of the Moose River Basin (Map 2543, map case; Vos 1982a). Over 400 drill holes, covering an area of about 20 km$^2$, have been drilled in the area (Telford et al. 1991). Two types of lignite are present: a woody variety, which contains compressed carbonized branches and trunks; and dull coal, which consists of more amorphous and homogenized material (Telford et al. 1991). A typical section through the lignite deposits is recorded in the OGS Onakawana B borehole (Sanderson and Telford 1985; Bezys 1989). In this hole 1 minor and 2 major lignite seams are present (between 37.8 and 47.7 m, 54.1 and 61 m, and 64 and 65.5 m). Interbedded with these lignite intervals are sandy clay, dark organic clays and quartz sand.

Between 1943 and 1946, approximately 2000 tonnes of lignite were stockpiled from the deposits in briquette form for an anticipated post-war energy shortage which did not materialize. Since that time, the Onakawana deposits have been repeatedly tested but not put into production. The volume of lignite present is estimated to be about 190 million tonnes (Telford and Long 1986).

# SUMMARY AND DISCUSSION

## Introduction

This chapter has reviewed the Paleozoic and Mesozoic stratigraphy of Ontario within the context of sequence stratigraphic concepts. This succession was subdivided into 10 depositional sequences (*see* Figure 20.3; Chart D, Map 2582, map case), defined as stratigraphic intervals of genetically related sedimentary rocks, commonly bounded by unconformities. The spatial and temporal variations in sediment type and distribution within these sequences, summarized from Chart D (Map 2582, map case) in Figure 20.48 and shown in Figures 2 to 11 on Map 2582 (map case), reflect the transgression and regression of basin-centred inland seas onto the North American craton during this time and indicate relationships between tectonics and sedimentation.

The 4 Paleozoic and Mesozoic sedimentary basins in and adjacent to Ontario (*see* Figure 20.2; Figure 1, Map 2582, map case) can be classified according to tectonic setting with characteristic morphology and sediment infilling. The ovoid Appalachian Basin is a foreland basin, formed by orogenic processes at the craton margin and filled mainly with orogen-derived clastic sediments. The remaining 3 basins, the Michigan, Hudson Bay and Moose River basins, are intracratonic basins, roughly circular and filled mainly with carbonate sediments, craton-derived clastic sediments and evaporites. The factors controlling formation of these latter basins are less well understood, but are probably linked to craton margin tectonics.

The processes of basin development and sedimentation are inevitably tied to tectonic processes and events. Tectonic controls are readily apparent at a large scale and close to orogenic belts. However, at a more detailed scale and farther out onto the craton, cause and effect relationships become less obvious and more complex, especially with respect to the chronostratigraphic correlation of sedimentary and erosional processes with tectonic events. Supply of clastic sediments from highlands created by plate margin tectonism is an example of a direct cause and effect relationship. Tectonic controls on intracratonic basin subsidence and arch uplift, as reflected in the deposition of sequences and the development of unconformities, are less obvious and are probably the consequence of a complex interplay of global, continental and basin-scale processes.

The following sections summarize the Paleozoic and Mesozoic history of Ontario, examine the tectonic controls on cratonic sedimentation as reflected by sequence stratigraphy, and discuss the implications for models of sedimentary basin evolution.

## Depositional History of Southern Ontario and the Lake Timiskaming Outlier (Western and Central St. Lawrence Platform)

The oldest Paleozoic sedimentary rocks in the province are craton-derived clastic rocks deposited in the Central and Western St. Lawrence Platform of southern Onta[illegible]ng the Early to Middle Cambrian (*see* Sequence 1; Fig[illegible] and Chart D, Map 2582, map case). During Sequence 1, the Michigan Basin was separated by the Algonquin Arch from the continental shelf, which was to become the Appalachian Basin in the southeast (Figure 2, Map 2582, map case). Craton-derived clastic sediments, succeeded by shallow water carbonate sediments, progressively onlapped the craton during Sequence 1. Middle Ordovician uplift, related to the onset of the Taconic Orogeny, resulted in the development of a major unconformity (Sauk–Tippecanoe boundary of Sloss 1963; *see* Figure 20.48), terminating Sequence 1 deposition.

During Sequence 2, early phases of the Taconic Orogeny along the eastern margin of the craton created a marginal trough, the incipient Appalachian Basin, which then began to receive orogen-derived sediments (Figure 3, Map 2582, map case). The Michigan Basin and Algonquin Arch became less distinct features during this sequence as a dominantly carbonate succession was deposited on a broad shelf covering this area.

Following Sequence 2, a major pulse of clastic detritus from the Taconic Orogeny filled the Appalachian Basin and spilled over into the Michigan Basin during Sequence 3 (Figure 4, Map 2582, map case). The major unconformity which caps this sequence may represent a glacio-eustatic sea level drop related to the continental glaciation of North Africa (Dennison 1976).

Orogen-derived clastic sedimentation continued during Sequence 4, with carbonate sediments becoming more prominent in the Michigan Basin. The Michigan Basin, Algonquin Arch and Appalachian Basin configuration in southern Ontario became firmly established by Sequence 4 (Figure 5, Map 2582, map case); however, lateral movements in the Algonquin Arch resulted in complex sedimentary depositional patterns and arch-centred unconformity development (e.g., Brett et al. 1990; Duke 1991).

Relative tectonic quiescence during Sequence 5 contributed to the development of a carbonate platform around the Michigan Basin and on the northwestern margin of the Appalachian Basin (Figure 6, Map 2582, map case). Basin to arch facies differentiation resulted from continued basin subsidence. Minor orogen-derived clastic sediments continued to be deposited from the Taconic Highlands during Sequence 5, but were largely restricted to the Appalachian Basin.

During Sequence 6, late Taconic or early Acadian orogenic activity (the Salinic Disturbance?) along the craton margin resulted in cratonic uplift, especially along the arches, with concomitant restricted basin circulation and evaporite deposition (Figure 7, Map 2582, map case). This orogenic activity also shed orogen-derived clastic sediments into the Appalachian Basin. Widespread cratonic uplift or eustatic sea level drop in the Early Devonian resulted in the termination of Sequence 6 deposition and development of a major unconformity (Tippecanoe–Kaskaskia boundary of Sloss (1963); *see* Figure 20.3).

Initiation of Sequence 7 was marked by the deposition of craton-derived clastic sediments in continental settings (Chart D, Map 2582, map case). Subsequently, as transgression proceeded, widespread carbonate deposition occurred in both the Appalachian and Michigan basins, with the Algonquin Arch playing a subordinate role in controlling facies distribution. Evaporite deposition in the Michigan Basin during the Middle Devonian indicates restriction of the basin by the Algonquin Arch during the latter part of this sequence (Figure 8 and Chart D, Map 2582, map case).

Major orogenic activity along the eastern continental margin, in the form of the Acadian Orogeny, commenced during Sequence 8. Vast amounts of orogen-derived clastic detritus were generated and deposited over all of southern Ontario and northward to the Hudson Platform (Figure 9, Map 2582, map case).

Orogen-derived clastic sediments shed from the Alleghanian Orogeny during Sequence 9 were deposited in the Appalachian and Michigan basins and over the Algonquin Arch (Figure 10, Map 2582, map case), although subsequent erosion removed these sediments from southern Ontario.

By the Mesozoic, collisional tectonism had essentially ceased along the eastern margin of North America. There is no record of sedimentation in southern Ontario during Sequence 10, however, some Mesozoic age craton-derived clastic sediments were deposited in the centre of the Michigan Basin (Figure 11, Map 2582, map case).

Extensional tectonism during the Mesozoic, related to the opening of the Atlantic Ocean, resulted in the local development of block faulting and emplacement of ultramafic and mafic dikes. The general relationship between rifting and dike emplacement is discussed in this volume, mainly with respect to Proterozoic events (Osmani, this volume).

## Depositional History of Northern Ontario (Hudson Platform)

Paleozoic sedimentation did not commence in the Hudson Platform until the late Middle Ordovician (*see* Sequence 2; Chart D, Map 2582, map case). Towards the end

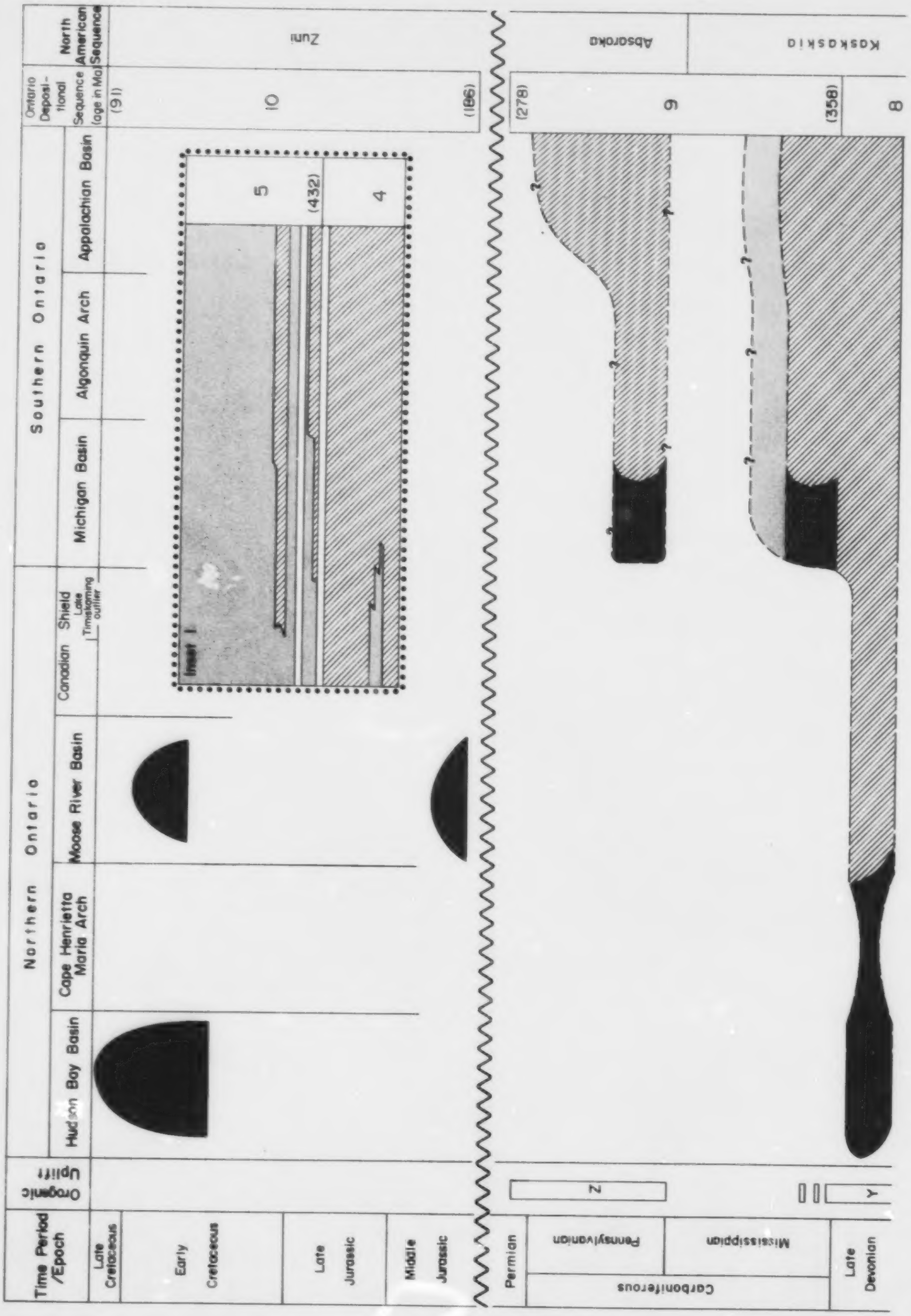

**Figure 20.48.** Spatial and temporal distribution of Paleozoic and Mesozoic depositional sequences (and their constituent sediment types) in Ontario; lithologic summary of Chart D (*see* Map 2582, map case).

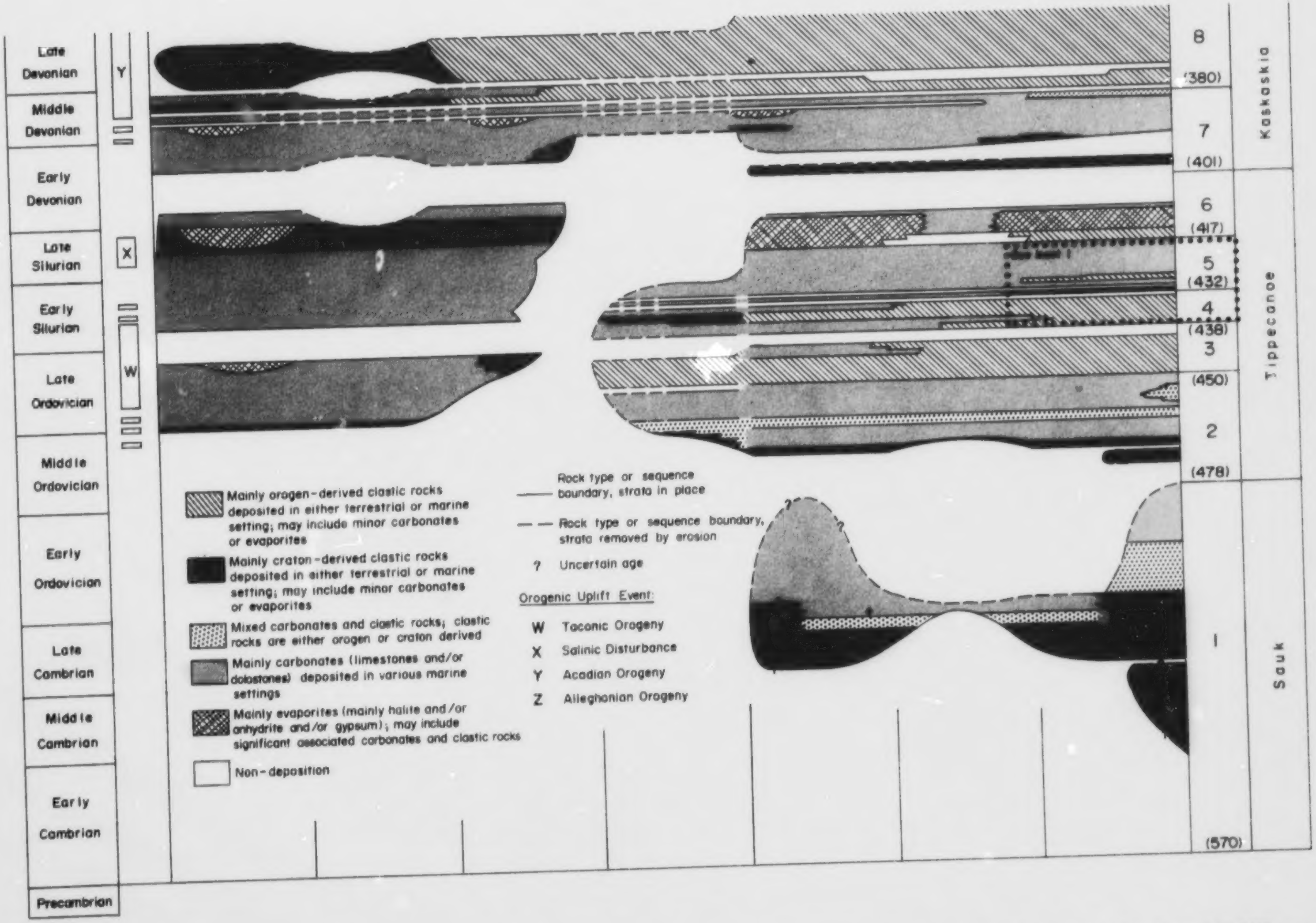

Figure 20.48. Spatial and temporal distribution of Paleozoic and Mesozoic depositional sequences (and their constituent sediment types) in Ontario; lithologic summary of Chart D (*see* Map 2582, map case).

of Sequence 2 time, craton-derived clastic sediments, succeeded by carbonates, were deposited in the Hudson Bay Basin and in the northern end of the Moose River Basin (Figure 3, Map 2582, map case).

By Sequence 3 time, the Moose River Basin was better defined, and uplift of the Cape Henrietta Maria Arch during the latter stages of that sequence resulted in a clearer differentiation of depositional facies in the 2 basins (Figure 4 and Chart D, Map 2582, map case). Although the orogen-derived clastic sediments deposited during Sequence 3 in southern Ontario did not extend into the Hudson Platform, craton-derived clastic sediments in the Moose River Basin and evaporite deposition in the Hudson Bay Basin indicate uplift of adjacent cratonic areas during the latter part of Sequence 3.

Sequence 4 deposition in the Hudson Platform, consisting mostly of carbonate sediments, did not commence until late in the sequence (Chart D, Map 2582, map case), perhaps in response to cratonic uplift following the Taconic Orogeny. The upper boundary of Sequence 4 in the Hudson Platform appears to be conformable, marking the transition into an environment conducive to reef development.

The marine transgression, which had begun in Sequence 4, was firmly established during Sequence 5, at which time major carbonate build-ups were developed along basin margins (Figure 6, Map 2582, map case). Carbonate build-ups also developed along the uplifted Central Hudson Bay Arch (*see* Figure 20.30). The Late Silurian uplift during Sequence 6 deposited craton-derived clastic sediments across the Hudson Platform and evaporites in the Hudson Bay Basin.

Following the widespread erosional unconformity between sequences 6 and 7 (Tippecanoe–Kaskaskia boundary of Sloss 1963; *see* Figure 20.48), the 2 basins were once again established. Although carbonate sedimentation dominated during Sequence 7 time, craton-derived clastic sediments were deposited in the Moose River Basin and basin-centred evaporites were deposited in both the Moose River and Hudson Bay basins near the close of this sequence.

The orogen-derived clastic sediments shed from the Acadian Orogeny during Sequence 8 time were deposited in the Moose River Basin, with their northern limit defined by the Cape Henrietta Maria Arch. Shallower water, craton-derived clastic sediments and minor carbonates were deposited in the Hudson Bay Basin during this time.

There is no record of sedimentation in the Hudson Platform during the time interval represented by Sequence 9 (Figure 10 and Chart D, Map 2582, map case). This interval of nondeposition or erosion may have been due to cratonic uplift related to the Alleghanian Orogeny.

The Moose River Basin and Hudson Bay Basin appear to have been separate entities during Sequence 10 time. Craton-derived clastic sediments were deposited in a terrestrial environment in the former and in a marine setting in the latter basin (Figure 11, Map 2582, map case).

## Paleozoic and Mesozoic Igneous Activity and Extensional Tectonism

A number of Paleozoic and Mesozoic mafic, ultramafic, kimberlitic, alkalic and carbonatite intrusions (including dikes, plugs, pipes, diatremes and complexes) have been indentified in Ontario. The tectonic setting, distribution and ages of these intrusions are summarized here from more detailed discussions presented in chapters by Sage (this volume) and Easton (this volume), and in Sequence 10 of this chapter. The general location and age ranges of these intrusions are plotted on Chart D, Map 2582 (map case).

Most of the Paleozoic and Mesozoic intrusions in Ontario are located within 1 of 2 major structural features in the province, the Ottawa–Bonnechere Graben and Lake Timiskaming graben or Fault System (*see* Figure 20.2). The Ottawa–Bonnechere Graben may be traced along the Ottawa River from Montreal to Mattawa, where it trends westward through Lake Nipissing to the North Shore of Lake Huron. The Lake Timiskaming graben extends northwestward, from the Lake Timiskaming–Kirkland Lake area, to the Moose River Basin where it intersects the Kapuskasing Structural Zone (*see* Sage, this volume), and southeastward towards Mattawa where it connects with the Ottawa–Bonnechere Graben. Some workers have suggested that the Ottawa–Bonnechere Graben is a branch of the St. Lawrence Graben System (Kumarapeli 1985). Adams and Basham (1989) suggested that the St. Lawrence Graben System may extend into Ontario, roughly paralleling the lower Great Lakes. In spite of the suggested linkage of some or all of these extensional tectonic features, there is uncertainty concerning the timing of initiation and reactivation of these features.

Some alkalic and carbonatite complexes in the Lake Nipissing area (Manitou Island and Newman Island complexes) have been dated at earliest Cambrian (*see* Sage, this volume; Easton, this volume). This suggests that activation (and initiation?) of the Ottawa–Bonnechere Graben was related to Neoproterozoic to early Paleozoic rifting along the St. Lawrence Graben System which led to the formation of the Iapetus Ocean (proto–Atlantic).

Easton (this volume) described mafic dikes, called the Rideau Swarm, which were intruded into Grenville rocks in the Kingston–Gananoque area (i.e., near the possible extension of the St. Lawrence Graben System). The age determinations for these dikes range from Middle Cambrian to Late Silurian (525 to 414 Ma; *see* Easton, this volume). The younger age determinations are problematic, occurring when the craton was within a dominantly compressional tectonic regime. The older determinations, however, are consistent with an Iapetus rifting event. The Neoproterozoic Grenville mafic dike swarm (*see* Easton, this volume) is also likely related to this rifting event.

Mid–Paleozoic age ranges are also reported for 2 alkalic-carbonatite complexes, the Brent Crater and the Hecla–Kilmer Alkalic Complex (*see* Sage, this volume). Millman et al. (1960) and Easton (this volume) argued that the Brent Crater, which occurs within the Ottawa–Bonnechere Graben, is an astrobleme and not an alkalic complex. Also, there is considerable uncertainty associated with the age determination for the Hecla–Kilmer Alkalic Complex (Bell and Blenkinsop 1980), which occurs in the Moose River Basin in the vicinity of the intersection of the Lake Timiskaming graben and the Kapuskasing Structural Zone.

Mesozoic extensional tectonism, related to the creation of the Atlantic Ocean, gave rise to widespread igneous activity in Ontario and Quebec. Alkalic and ultramafic intrusions and dikes are known from the Monteregion Hills of southwestern Quebec (Manspeizer et al. 1989), and along the Ottawa–Bonnechere Graben and Lake Timiskaming graben to the Moose River Basin (*see* Sage, this volume). Jurassic ultramafic dikes cut Ordovician strata at Varty Lake and Picton, in the Belleville–Kingston area of southeastern Ontario (*see* Sequence 10; Barnett et al. 1984).

Mesozoic kimberlitic dikes which occur in the Lake Timiskaming–Kirkland Lake area (described briefly by Sage, this volume), have been the target of diamond exploration (Brummer 1978). Paleozoic strata in the Moose River Basin are cut by scattered 150 to 180 Ma (Janse et al. 1986) mafic and ultramafic alkalic diatremes of kimberlitic affinity. Occurrences are reported north of the Martinson Carbonatite Complex (plotted on Map 2543, map case) and in the Kapuskasing Structural Zone at Coral Rapids and Sextant Rapids on the Abitibi River (*see* Sage, this volume; Sanford and Norris 1975).

## Sequence Stratigraphy

The application of sequence stratigraphic concepts to the Paleozoic and Mesozoic succession of Ontario has resulted in the identification of 10 depositional sequences (Chart D, Map 2582, map case; *see* Figures 20.3 and 20.48) and has provided a framework within which to examine tectonic influences on cratonic sedimentation. There are, however, a number of limitations to this application, and these are discussed below.

As previously mentioned, few of the 10 depositional sequences in Ontario correspond directly with Sloss's (1963) continent-wide first order sequences (*see* Figures 20.3 and 20.48). However, most of Sloss's sequence boundaries can be correlated with sequence-bounding unconformities in Ontario (within the resolution of biostratigraphic dating); the boundary between Sloss's (1963) Kaskaskia and Absaroka sequences lies within Sequence 9 and is unrecognized in Ontario because of the lack of preserved strata. The remaining sequence boundaries in Ontario must then reflect more local, smaller- scale (i.e., eastern North America or sedimentary basin) controls.

Additionally, there appears to be an inconsistency of scale among the Ontario sequences. For instance, Sequence 1 corresponds to the entire Sauk Sequence, whereas, the Tippecanoe Sequence consists of sequences 2 through 6 (*see* Figure 20.48). This inconsistency may be due, in part, to variabilities in exposure and preservation of the rock record; Sequence 1 is less well exposed than sequences 2 through 5 and much of Sequence 1 was removed during pre–Sequence 2 erosion. However, sequences 2 through 5 may also reflect the more tectonically active interval during the Taconic and Acadian orogenies as compared with the relatively stable pre–Taconic Sequence 1 time.

The data used in constructing this sequence stratigraphic framework are subject to some skewing, based on the availability of data (or lack thereof); that is, some stratigraphic intervals are better known than others. In a geographic sense, the entire Paleozoic and Mesozoic succession in the Hudson Platform is not well known due to its remoteness, very poor surface exposure and limited subsurface data. This makes interbasinal correlation of sequence boundaries difficult and in some cases tenuous. For instance, the boundary between sequences 4 and 5 is a regional angular unconformity in southern Ontario, and yet appears to be conformable in northern Ontario.

Ideally, the sedimentary record of a depositional sequence should reflect transgression followed by regression. The patterns of sediment deposition illustrated in Chart D, Map 2582 (map case) are far more complex and the reasons for this are numerous. Various depositional settings will reflect transgression and regression differently, as will position within a given setting or basin. Preservation in the rock record of an "ideal sequence" is unlikely, due to such factors as rate of relative sea level change and subsequent erosional events. The main factor complicating the patterns of sedimentary deposition may be the superposition of effects from processes acting on different scales (e.g., eustatic sea level drop caused by worldwide glaciation combined with basin subsidence caused by overthrust loading).

Another significant factor complicating interbasin sequence correlation in Ontario is the variation in tectonic setting across the province. Undoubtedly, the responses to plate margin tectonism, such as basin subsidence or arch uplift, vary in magnitude and timing across the craton. Thus, significant signatures (i.e., major unconformities) in the sedimentary record along the margin of a foreland basin may correspond to more subtle (and younger?) signatures preserved in the record of a distant intracratonic basin.

The constraints on the application of sequence stratigraphic concepts listed above have necessitated the usage of a broader definition of depositional sequence or sequence in this chapter than that proposed by previous workers (e.g., Mitchum et al. 1977). The definition used in this chapter is: a depositional sequence is a stratigraphic interval or package of genetically related sedimentary rocks which is commonly bounded by unconformities. This broader definition allows sequence boundaries to be drawn at significant changes in sediment type or pattern (e.g., between sequences 2 and 3) which may represent a more subtle response to tectonic activity.

## Tectonics, Sedimentation and Basin Development

Processes effecting cratonic sedimentation range from world-scale eustatic sea level fluctuations to continent-scale tectonic activity (e.g., orogenies) to basin-scale tectonics (e.g., arch uplift). These processes combine to produce a complex record of depositional sequences and

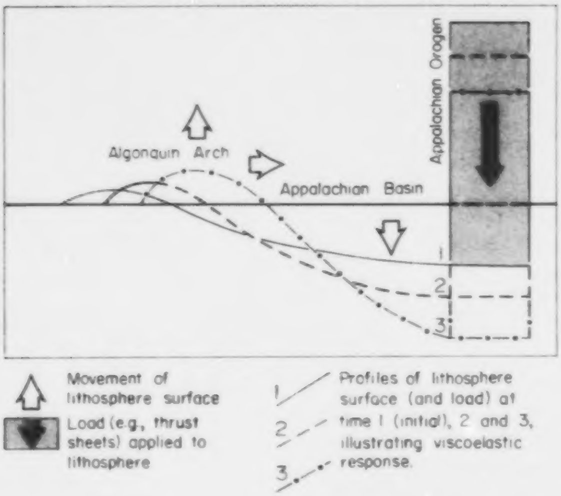
Appalachian Orogen
Algonquin Arch
Appalachian Basin
1
2
3
Movement of lithosphere surface
Load (e.g., thrust sheets) applied to lithosphere
1
2
3
Profiles of lithosphere surface (and load) at time 1 (initial), 2 and 3, illustrating viscoelastic response.

**Figure 20.49.** Schematic cross section illustrating lithospheric flexure model as applied to the Appalachian Basin and Algonquin Arch (*modified from* Quinlan and Beaumont 1984).

unconformities of varying scale and distribution. Examination of the sediment deposition patterns illustrated in Chart D and Figures 2 to 11, Map 2582 (map case) and summarized in Figure 20.48, reveals evidence of both intracratonic and craton-margin tectonic controls on sedimentation.

The 3 main Paleozoic collisional events, the Taconic, Acadian and Alleghanian orogenies, which occurred in the Appalachian Orogen on the eastern margin of North America, produced major pulses of orogen-derived clastic sediments during sequences 3 and 4, 8, and 9, respectively. Smaller clastic pulses (e.g., sequences 2, 5 and 6) suggest that orogenic activity may have continued, to a lesser degree, between the main orogenies. Periods of relative tectonic quiescence are indicated by intervals of widespread carbonate deposition (e.g., sequences 2, 5 and 7).

The Appalachian Basin and its dominantly clastic sediment content originated as a direct consequence of craton-margin tectonism (Quinlan and Beaumont 1984). Collisional tectonics to the south and east caused emplacement of thrust sheets, which initiated crustal loading, lithospheric downwarping, subsidence and formation of a foreland basin (Figure 20.49). Subsequent erosion of the highlands supplied clastic debris to the basin, probably causing further subsidence. The Algonquin Arch is postulated to be a peripheral bulge (*see* Figure 20.49), formed cratonward of the foreland basin, in response to lithospheric flexure (Quinlan and Beaumont 1984). The arch acted, in part, to confine orogen-derived sedimentation to the foreland basin and, as Quinlan and Beaumont (1984) suggested, may have rebounded during periods of tectonic quiescence, resulting in arch-centred erosion and unconformity generation.

Cratonic tectonism, such as intracratonic basin-centred subsidence and arch-centred uplift, is evidenced in the depositional patterns of Figure 20.48, Chart D and Figures 2 to 11 (Map 2582, map case), by:

1. post-depositional erosion and the lack of deposition over the shield area and the arches (not indicated in Chart D, Map 2582, because unit thicknesses are not plotted)
2. the partial restriction of orogen-derived clastic sediments to areas south and east of the Algonquin Arch
3. the basin-centred distribution of some sediments (e.g., evaporites) and sedimentary facies

The dominant type of clastic sediment in intracratonic basins is craton derived, generated by erosion of older sedimentary rocks and shield material from uplifted adjacent arch areas. Evaporite deposition resulted from arch uplift (and possibly basin uplift), which caused restriction of water circulation and promoted evaporation. Basin-centred subsidence can influence carbonate deposition by accentuating depth dependant facies development. This was especially evident during Sequence 5 in the Michigan Basin, when the Upper Silurian barrier, pinnacle and patch reef belts developed concentrically around the basin (*see* Figure 20.23).

The location of intracratonic basins, distant from orogenic belts, makes it more difficult to postulate a potential driving mechanism for their subsidence. In addition, there is disagreement among workers as to the commonality of crustal characteristics beneath intracratonic basins and the synchroneity of basin subsidence histories (*see* Williams, Stott, Thurston et al., this volume).

Numerous models have been proposed to explain the origin and evolution of sedimentary basins (e.g., Sloss and Speed 1974; McKenzie 1978; Klein and Hsui 1987). Models pertaining to intracratonic basins have been reviewed by Quinlan (1987) and are discussed by Williams, Stott, Thurston et al. (this volume). Perhaps the most plausible of these models involves transmission, through the crust, of stresses from collisional tectonic activity at the continental plate margin (e.g., Sandford et al. 1985; Cloetingh 1988; Howell and van der Pluijm 1990). Thus, arch uplift and basin subsidence on the craton would be accelerated during, or sometime after, collisional tectonic events at the plate margin. More detailed biostratigraphic work is required, especially in interior basins such as the Hudson Bay and Moose River basins, to better correlate the timing of cratonic tectonic activity (i.e., uplift and unconformity development or subsidence and sedimentation) with orogenic tectonic activity.

## Faults and the Fracture-Framework Model

Intracratonic tectonic activity may have also had a more localized effect on sedimentation. Sanford et al. (1985) proposed that the relative vertical movements of basins and arches were accommodated by brittle fracturing and faulting of the Precambrian basement and overlying sedimentary rocks, resulting in the development of a fracture-framework or network of fault blocks (*see* Figure 20.29). These authors further postulated that these fault blocks controlled local sediment distribution patterns and that the faults and fractures acted as conduits for fluid flow (both water and hydrocarbons).

Sanford et al. (1985) suggested that during the Silurian (*see* Sequence 5), reef growth was favoured on the upthrown edge of the fault blocks (i.e., closer to the photic zone) and that this resulted in the development of linear reef trends. These authors also suggested that post-depositional dissolution of Salina Formation (*see* Sequence 6) salt beds along linear (i.e., fracture-fault) trends resulted in depositional thickening of the overlying Bass Islands Formation. This mechanism may have also been active during the deposition of Middle Devonian (*see* Sequence 7) carbonate sediments and in the later development of structures in these strata which would form hydrocarbon reservoirs. The fractures and faults that cut the Middle Ordovician carbonate succession (*see* Sequence 2) were conduits for dolomitizing fluids. Dolomitization along these conduits enhanced the porosity of the host rocks and created some of the most significant hydrocarbon reservoirs in Ontario (Carter 1990).

Faults cutting or underlying Paleozoic and Mesozoic strata in Ontario are shown on the accompanying bedrock geology and tectonic assemblage maps (Maps 2541, 2543, 2544, 2575, 2577 and 2578, map case). These faults fall into 2 categories: those exposed in surface outcrops exhibiting a demonstrable vertical offset of Paleozoic or Mesozoic strata (e.g., eastern Ontario; Williams 1991); and those demonstrably offsetting Paleozoic strata in the subsurface with no surface expression (e.g., southwestern Ontario; Brigham 1971). The latter type are only identified in southwestern Ontario (Maps 2544 and 2578, map case) because of the relative abundance of subsurface information afforded by hydrocarbon exploration activity there. Both of these fault types commonly involve the Precambrian basement and, therefore, are generally traceable on aeromagnetic maps (e.g., Map 2587, map case). In this case, geophysical responses indicate basement structures which may have been reactivated to form faults in the overlying Paleozoic strata. In some cases, faults which have been identified in the Precambrian basement by geophysical methods have no known Paleozoic or Mesozoic displacement.

Although faults have been identified cutting the Paleozoic strata in southern Ontario, an extensive fracture-framework, such as that conceptualized by Sanford et al. (1985) and shown in Figure 20.29, has not been recognized. This can be attributed, in part, to the extensive cover of relatively young Quaternary sediments over most of the province. Additionally, one must consider the intensity of block faulting in relation to driving tectonic forces. Creation of fractures and faults and their subsequent reactivation presumably occurred during compressional or extensional tectonism at the plate margin. Since the Mesozoic, the eastern margin of North America has been tectonically passive, and perhaps reactivation of previously established faults and fractures has diminished since then. Much work remains to document the existence and extent of such a fracture-framework network and to determine how active it has been during the Phanerozoic.

Recent investigations have been undertaken to determine the origin of neotectonic (i.e., post-glacial) features in southern Ontario (e.g., Gorrell 1988; McFall and Allam 1989). There is some evidence that recent seismic activity and neotectonic structural features are related to reactivation of structures which cut Paleozoic and Precambrian strata (Adams and Basham 1989; McFall and Allam 1989, in prep.; Fakundiny, Myers et al. 1978; Fakundiny, Pferd et al. 1978).

## Concluding Remarks

This chapter has demonstrated how tectonic forces affected deposition of sedimentary rocks in Ontario. In controlling sedimentation, tectonic forces also controlled the origin and distribution of geological resources such as salt, gypsum, oil and natural gas, groundwater, aggregate, building stone, chemical stone and shale-clay, among others. Clearly, a better understanding of the Paleozoic and Mesozoic history of the province would increase the efficient exploration and development of our geological resources in an environmentally and economically sound manner.

Although a geological history is well established for certain intervals of the Paleozoic and Mesozoic in certain areas of the province, large gaps exist in our understanding of these rocks, particularly in certain stratigraphic intervals and geographic regions. Much more basic geological data is required for the entire Paleozoic and Mesozoic succession in the Hudson Platform and for covered parts of the succession in southern Ontario. Future data acquisition must include subsurface investigations, given the generally extensive Quaternary cover over the Paleozoic and Mesozoic basins. This chapter has demonstrated the necessity for more detailed biostratigraphic work, in order to provide a more accurate chronostratigraphic framework which is necessary to advance our understanding of the relationships between tectonism and the sedimentary record. Future investigations must incorporate and evaluate new models and concepts such as sequence stratigraphy, basin evolution and the fracture-framework. These are unifying concepts which help link crustal processes to the rock record and ultimately to our environment.

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# Quaternary

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# Chapter 21

# Quaternary Geology of Ontario

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## Abstract

**Twenty thousand years ago, Ontario was completely covered by ice of the Laurentide Ice Sheet. Part of a large continental glacier complex, this ice sheet covered much of Canada and extended into the northern United States. This was not the first time that Ontario was glaciated. Great ice sheets covered parts of Ontario several times during the Quaternary Period as well as during the Precambrian.**

**Glaciations during the Quaternary Period have played a major role in shaping and creating the landscape of Ontario. The deposits and effects of these glaciations are widespread. Till is the most widespread deposit left by the Laurentide Ice Sheet. As a result of the incorporation of a variety of bedrock rock types and existing sediments from the preglacial landscape of Ontario, 3 main end members of till were produced: sandy tills derived from the erosion of Precambrian rocks; silty tills derived from the erosion of carbonate rocks; and clayey tills derived from the incorporation or deformation of fine-grained glaciolacustrine sediments that were deposited in front of the glacier in local or regional areas of ponding.**

**Within the Great Lakes–St. Lawrence Lowlands, numerous till sheets were deposited during the advance and retreat of the Laurentide Ice Sheet margin. Ice flow was controlled by the broad topographic depressions of the Great Lakes basins. Lobes of ice extending beyond the main body of the ice sheet developed in these basins and acted at times independently in response to local conditions at the base of the glacier rather than, or in addition to, climatic change. An account of these ice-marginal oscillations is given and the physical characteristics of some of the till sheets is summarized.**

**Considerable volumes of meltwater were generated by, and discharged from, the glaciers that once occupied Ontario. Large amounts of glacial debris were transported by the meltwater and deposited as stratified sediments under the glacier in conduits and cavities and along the ice margin (glaciofluvial ice-contact deposits), and beyond the ice margin in rivers and streams (glaciofluvial outwash deposits), lakes (glaciolacustrine and lacustrine deposits) and seas (glaciomarine and marine deposits). Stratified sediments deposited by this meltwater are extensive and are the predominant surficial sediment throughout a large part of Ontario. The series of proglacial ice-contact lakes that formed and disappeared with the fluctuations of the ice margin during deglaciation are described.**

**Some of the properties and characteristics of the deposits left behind by the Laurentide Ice Sheet are introduced and some of their applied aspects discussed. The Quaternary Geology of Ontario maps accompanying the volume display the major features of the Quaternary geology of the province.**

## INTRODUCTION

### Prelude

This chapter reviews the Quaternary geology of the province of Ontario. The focus of the chapter is on the record of Quaternary glaciations which have shaped the landscape of the province. The chapter provides a brief review of glacially derived landforms and sediment types, the Quaternary stratigraphy of Ontario, and an overview of the complex history of the proglacial lakes which evolved into the Great Lakes.

Twenty thousand years ago, Ontario was completely covered by the Laurentide Ice Sheet. Part of a large continental glacier complex, this ice sheet covered much of Canada and extended into the northern states of the United States (Figure 21.1). This was not the first time that Ontario was glaciated. Great ice sheets covered parts of Ontario several times during the Quaternary Period as well as during the Precambrian, some 2.4 billion years ago (*see* Bennett and Dressler, this volume).

Quaternary glaciations have played a major role in shaping and creating the landscape of Ontario. They removed pre-existing sediment and soils, eroded and smoothed bedrock surfaces, formed landforms such as moraines and eskers that are peculiar to glaciers, and left unconsolidated deposits of till, gravel, sand, silt and clay.

The deposits and landforms left by the glaciers have played an important role in the history and development of the province of Ontario. Many of the early explorers' routes followed ancient paths of glacial meltwater. The lakes, rivers, streams and springs, around which the native peoples and early settlers congregated, originated during and following the last glaciation. Our fertile soils are young, having been developed on glacial deposits. The origin of the glacial sediments, whether deposited directly by the ice in streams emanating from the ice or in lakes supported by the ice, determines their properties and consequently what types of soils will form and what types of plants will prosper.

Knowledge of the distribution of the sediments deposited by the glaciers and of the glacial history of Ontario is important to plan wisely the use of this province's natural resources. Whether it be a scenic landscape for use as green space, or a deposit of sand and gravel for use in concrete in our homes and highways, it is essential to know what our resources are and how they can be managed and used to our best advantage.

The Quaternary deposits of Ontario also provide a record of major changes in climate and global conditions over the past 140 000 years. This record, although incomplete, contains evidence of the terrestrial system's response to major shifts in climate and provides analogies for future changes that may occur in Ontario as a result of climatic change.

This report amplifies and explains the accompanying 4 map sheets (*see* Maps 2553, 2554, 2555 and 2556, map case) of the Quaternary Geology of Ontario. The first section serves as an introduction to the Quaternary Period, the Laurentide Ice Sheet and the types of deposits and landforms left by continental glaciers. In the second section, the deposits and landforms left by the Laurentide Ice Sheet in Ontario are discussed and the glacial history of Ontario is summarized. The third section describes some of the applied aspects of sediments the glaciers left behind.

There are many differing views on the interpretation of the glacial history of Ontario. Several summaries have been published that include discussions of this history. The works of Prest (1970) and Shilts et al. (1987), and various authors in Fulton (1984a, 1989a) and Fulton and Andrews (1987) should be consulted for alternate perspectives on this exciting period of the earth's history.

## The Quaternary Period

The Quaternary Period is the youngest period of the Earth's history. Beginning about 1.8 million years ago, it is generally considered a time when the Earth's climate was cool and when large parts of North America and Europe were intermittently covered by large continental ice sheets. The Quaternary Period has been subdivided into the Pleistocene and Holocene epochs. The Pleistocene Epoch, often referred to as the Great Ice Age, is the period of time during which the growth and decay of several continental-scale ice sheets occurred. The Holocene Epoch, also referred to as the Recent, includes our present postglacial time beginning approximately 10 000 years ago, up to and including the present.

The geological record of the Quaternary Period in Ontario is quite extensive; however, it is very incomplete. This is primarily because Ontario is situated in the centre of the North American land mass, a terrestrial location, where erosional processes are constantly removing parts of the geological record. More complete records of the Quaternary Period have been found; in the deep ocean basins of the Atlantic, Pacific and Indian oceans (Hays et al. 1976; Imbrie et al. 1984; Labeyrie et al. 1987; Martinson et al. 1987; and many others); in ice cores from the Greenland and Antarctic ice sheets (Dansgaard et al. 1982; Lorius et al. 1985; Jouzel et al. 1989); and in thick terrestrial loess sequences in Czechoslovakia, China and central Asia (Kukla 1975; Liu et al. 1985; Smalley 1987).

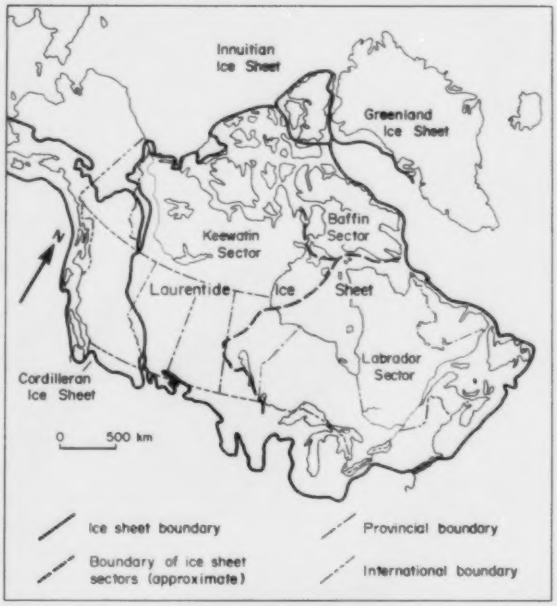


**Figure 21.1.** The North American glacier complex and the distribution of the Laurentide Ice Sheet (*modified from* Fulton 1989a).

The results obtained from studies of the faunal, floral and isotopic composition of the deep-sea cores provide a measuring stick against which the incomplete terrestrial record can be compared. Variations in the oxygen isotope composition of the calcareous shells of foraminifera found in these deep-sea cores have been used to subdivide the Quaternary Period into 63 oxygen isotope stages. These stages are numbered from 1, the youngest, which represents the Holocene, to 63, which existed some 1.8 million years (Ma) ago (Figure 21.2). Variations in the oxygen isotope, $\delta^{18}O$, content in the cores studied are believed to reflect changes in the volume of continental ice sheets through time. In general, the lighter isotope of oxygen , $^{16}O$, is preferentially evaporated from the oceans and transferred to land as precipitation. During continental glaciation, the lighter isotope of oxygen is stored in the ice sheets, resulting in greater concentrations of the heavier oxygen isotope , $^{18}O$, in the ocean waters. The isotopic composition of the foraminifera which live in the ocean waters reflects the isotopic composition of the ocean waters they live in.

Temperature change through the Quaternary Period may be determined by analyzing the variations in the content of deuterium, the heavy isotope of hydrogen, in ice cores from Antarctica and Greenland (Jouzel et al. 1989). Temperature curves derived from ice cores dating back to 160 000 years ago compare favourably with the oxygen isotope curves obtained from deep-sea cores (Figure 21.3). It may be possible to use the oxygen isotope curves as an approximation of temperature change, as well as ice volume change, throughout the entire Quaternary Period, at least until longer and older ice-core records are obtained. The ice cores also provide valuable information on changes in atmospheric composition through time.

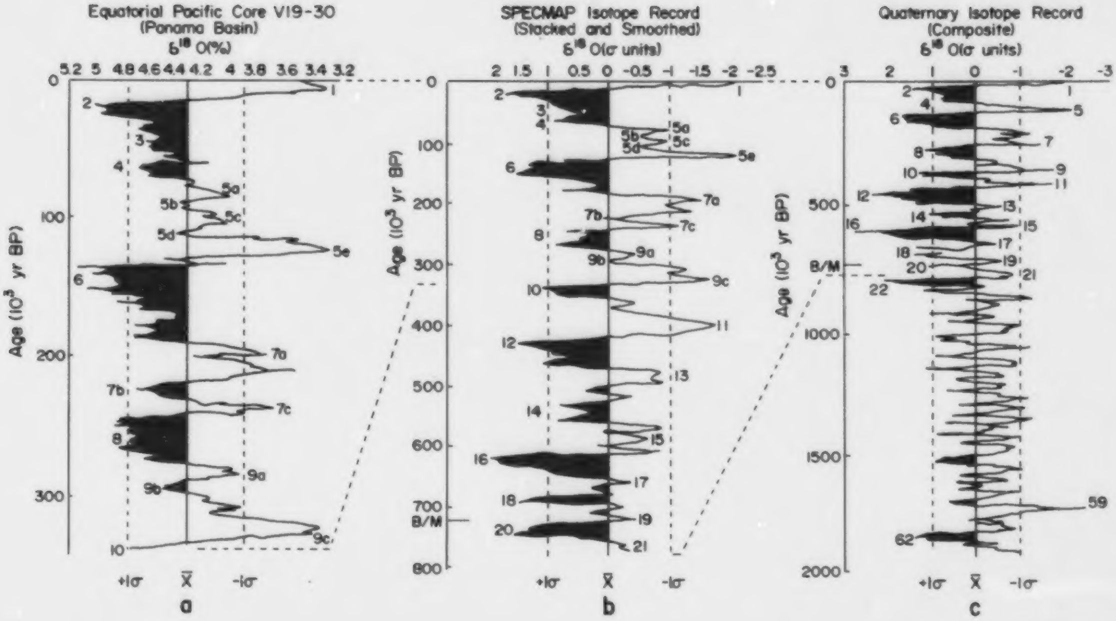


**Figure 21.2.** A composite marine oxygen isotope record of the Quaternary Period (*modified from* Porter 1989). Numbered isotope stages and substages are indicated and the level of the Brunhes/Matuyama magnetic reversal (B/M) is shown on the 2 longest records. **a)** equatorial Pacific core V19-30 derived from benthic foraminiferal sampled at ca. 1050-year intervals (data *modified from* Shackelton and Pisias 1985, chronology based on Martinson et al. 1987); **b)** stacked and smoothed SPECMAP record from 5 low- and middle-latitude deep-sea cores (Imbrie et al. 1984); and **c)** composite curve derived from 4 normalized isotope records (Williams et al. 1988).

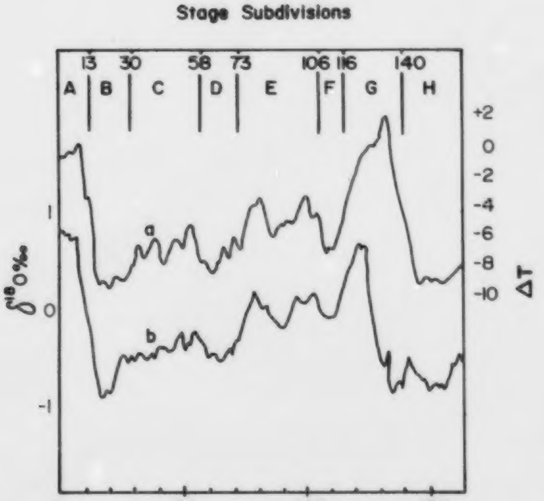


**Figure 21.3.** Comparison of the isotope temperature record (curve a), derived from the deuterium content in Vostock ice-core records (*modified from* Jouzel et al. 1987) and the marine oxygen isotope record (curve b; *modified from* Martinson et al. 1987); *modified from* Jouzel et al. 1989.

Quaternary deposits in Ontario appear to represent only the last 6 oxygen isotope stages or approximately the last 190 000 years. However, this includes 2 main glacial stages, the Illinoian and the Wisconsinan, the interglacial Sangamonian Stage, and the Holocene or Recent Epoch.

The Illinoian Stage corresponds to oxygen isotope stage 6 in the deep-sea cores. It represents a relatively cold period when most, if not all, of Ontario was covered by a large continental glacier. Based on ice-core data, the temperature shift associated with this glaciation, and with the younger Late Wisconsinan glaciation, was on the order of 9°C colder than present (Jouzel et al. 1989).

The Sangamonian Interglacial Stage, corresponding to oxygen isotope stage 5e approximately 115 000 to 135 000 years ago, is a period of time when the climate was as warm or warmer than it is today. This is contrary to Fulton (1984b, 1989b), but in agreement with Karrow (1989). As defined here, direct correlation can be made between the terrestrial deposits which contain floral and faunal remains indicating warm climate, and the oceanic deep-sea cores. The Sangamonian Interglacial Stage, as defined by Fulton (1989b), includes all of oxygen isotope stage 5, or times when the volumes of ice on the continents were substantial and temperatures were much cooler than today (3°C to 7°C cooler, Jouzel et al. 1989). With this definition, it would be difficult to separate nonglacial sediments deposited during interglacial cool periods from interstadial nonglacial sediments deposited during glacial periods. This could be only done by absolute dating techniques, which at present are severely wanting. Radiocarbon dating, the only proven absolute dating technique applicable to terrestrial deposits found in Ontario, is only reliable back to about 40 000 years ago.

The limited definition suggested here for the Sangamonian satisfies most of the concerns of Fulton (1984b). One of his concerns, in particular, is that an interglacial stage be defined on a global rather than a local scale. The Sangamonian is correlated to the oceanic oxygen isotope stage 5e and the Vostock ice-core record stage G (Figure 21.4). In addition, this definition is useful on the local scale where nonglacial sediments can be distinguished as interglacial or interstadial on the basis of their floral and faunal remains.

The Wisconsinan glaciation began approximately 115 000 years before present (BP) following the warm Sangamonian Interglacial Stage. The end of the Wisconsinan glaciation has been placed at 10 000 years ago, the beginning of the Holocene Epoch. The Wisconsinan Stage is commonly divided into Early, Middle and Late Wisconsinan; each with several subdivisions (*see* Figure 21.4). The Early Wisconsinan corresponds to the initial cooling and the inception and growth of the Laurentide Ice Sheet. The Middle Wisconsinan was slightly warmer than the Early Wisconsinan. The southern part of Ontario was essentially ice-free during this period of continental glaciation. During the Late or Classical Wisconsinan, which reached its maximum about 20 000 years ago, all of Ontario was once again covered by ice. By 15 000 years BP, the front of the Laurentide Ice Sheet began its final recession northward. By 10 000 years BP, deglaciation had progressed far enough northward for some to consider that the Holocene Epoch had begun. A large part of northern and central Ontario, however, was still ice covered. Not until after 8000 years BP had the Laurentide Ice Sheet left Ontario, and even today isostatic rebound continues. This is the rising of the Earth's crust in response to the removal of the mass of the continental ice sheets (Clark and Persoage 1970).

## The Laurentide Ice Sheet

The Laurentide Ice Sheet is the name given to the continental glacier that occupied about 80% of Canada during the Wisconsinan Stage (*see* Figure 21.1; Prest 1984; Fulton and

| Oxygen Isotope Stage + | Vostock Ice Core Stage * | Approximate Age (ka) | Classification following Dreimanis and Karrow (1972) | | Approximate Radiocarbon Age |
|---|---|---|---|---|---|
| 1 | A | | Holocene | | |
| | | 10 | | | |
| | | | | Port Huron Stade | 13 |
| | | | | Mackinaw Interstade | 13.4 |
| 2 | B | | Late Wisconsinan | Port Bruce Stade | 14.8 |
| | | | | Erie Interstade | 15.5 |
| | | | | Nissouri Stade | 20 |
| | | 30 | | | |
| | | | | Plum Point Interstade | 22 + |
| 3 | C | | Middle Wisconsinan | Cherrytree Stade | |
| | | | | Port Talbot Interstade | >40 |
| | | 60 | | | |
| 4 | D | | | Guildwood Stade | |
| | | 75 | | | |
| 5a, b, c | E | | Early Wisconsinan | St. Pierre Interstade | 75 |
| | | 105 | | | |
| 5d | F | | | Nicolet Stade | |
| | | 115 | | | |
| 5e | G | | Sangamonian | | |
| | | 135 | | | |
| 6 | H | | Illinoian | | |
| | | 190 | | | |

* from Jouzel et al. 1989

+ from Porter 1989

**Figure 21.4.** Subdivision of the Quaternary as related to Ontario deposits.

Prest 1987). It was part of a glacier complex in North America, which also included the Greenland and Cordilleran ice sheets as well as several smaller glaciers.

The Laurentide Ice Sheet covered all of Ontario when it was at its maximum southern extent, about 20 000 years ago. The ice sheet extended well into the northern United States, south of lat. 40°N (*see* Figure 21.1). The ice sheet extended westward almost to the Rocky Mountains, eastward to the continental shelf off the coast of the Atlantic provinces and northward to the Queen Elizabeth Islands (Fulton and Prest 1987).

Reconstructions of the Laurentide Ice Sheet have been quite varied. Originally, the Laurentide Ice Sheet was thought to be a complex of essentially independent glaciers with sources in Keewatin, Quebec–Labrador, and in northwestern Ontario (the Patrician glacier; Tyrrell 1898, 1913). Flint (1943), however, considered it to be a single ice sheet centred on Hudson Bay. He suggested that it originated in the highlands of Quebec–Labrador as valley glaciers which coalesced to form piedmont glaciers that eventually flowed westward into Hudson Bay. Eventually, a single dome formed, centred over Hudson Bay (Flint 1943).

Ives (1957; Ives et al. 1975) suggested the concept of "instantaneous glacierization" as an alternative to explain the mechanism of growth of the Laurentide Ice Sheet. Ives suggested that snow remaining on plateau areas of the Canadian Shield altered atmospheric circulation such that precipitation increased and snowlines lowered. This resulted in large expanses of shield plateaus, such as in Keewatin, Baffin Island and Quebec–Labrador, being covered by glaciers. Ives (1957), however, continued to support the idea of a solitary ice centre (over Hudson Bay) for the Laurentide Ice Sheet.

Prest (1957), however, based on field data and the pattern of streamlined landforms, displayed 2 main centres of ice flow in his reconstruction of the Laurentide Ice Sheet. The work of Shilts (1978, 1980) on the composition of the tills along the west coast of Hudson Bay demonstrated that ice flowed into Hudson Bay from Keewatin during the Late Wisconsinan. In addition, Hardy (1977) suggested that there is no evidence for ice flowing eastward out of Hudson Bay on the east side of Hudson Bay. Although seriously questioned, the concept of a single-domed ice sheet flowing radially out of Hudson Bay is still considered, particularly in numerical models of ice sheet growth and decay (e.g., Denton and Hughes 1981).

The Laurentide Ice Sheet has been divided into 3 sectors; the Labrador, the Keewatin and the Baffin sectors (Prest 1984; Fulton 1989b). Only the Labrador and Keewatin sectors directly affected Ontario.

Ice accumulation of the Labrador Sector began in the interior uplands of Labrador and Quebec. The ice flowed radially outward from the interior uplands and eventually merged with the Baffin Sector to the north and with the Keewatin Sector to the west. Ice also extended southward into the Great Lakes–St. Lawrence Lowlands and onto the continental shelf off the Atlantic coast. Labrador Sector ice extended westward well into the Prairies, but was later displaced by ice of the Keewatin Sector (Prest 1963, 1984; Prest and Nielsen 1987).

The Keewatin Sector of the Laurentide Ice Sheet originated in the district of Keewatin. The plateaus in the district of Keewatin are lower and much farther removed from sources of moisture than are the high plateaus adjacent to the North Atlantic Ocean which served as ice centres for the Labrador and Baffin sectors. This may have caused a delay in the formation of Keewatin Sector ice, allowing the Labrador Sector ice to spread westward and northwestward across northern Ontario and into the southern Prairies (Figure 21.5; Vincent and Prest 1987). Keewatin Sector ice expanded radially to eventually cover the southern Arctic islands, reach the eastern side of the Cordilleran mountain chain and extend southward into the north-central United States. Its flow was deflected to the northeast and to the southwest when it coalesced with ice of the Labrador Sector (Prest 1984). In Ontario, only the most western parts of the province were directly affected by the Keewatin Sector.

## Glacial Deposits and Landforms

### INTRODUCTION

A glacier moves by internal deformation, basal sliding and subsole deformation (Sugden and John 1976; Shaw 1985). Which one of these processes is dominant at a given location depends on the glacier's thermal regime and the composition and strength of the underlying material. If the glacier is frozen to its bed, no basal sliding can occur and erosion by abrasion is negligible; movement is primarily by internal deformation (Shaw 1985). With melting at the bed, basal sliding and erosion by abrasion are significant. Subglacial meltwater, an important erosional agent, is present as well. When a glacier overrides soft unconsolidated sediments, it is possible for its flow to be explained solely by subsole deformation within the underlying sediments.

Glaciers, in particular those of continental scale, usually have complex thermal regimes. In general, however, it is common for glaciers to be warm, temperate, or wet based in their accumulation zones (areas of net growth of ice) and cold or having cold patches in their ablation zones (areas of net ablation or melting). Glaciers with complex thermal regimes are capable of entraining (incorporating) large quantities of debris (Shaw 1985). Faulting and folding as a result of differential flow, particularly in the lower part of the glacier, may help in the mixing and homogenization of debris carried at the base of the ice. In continental glaciers, most erosion and incorporation of material into the glacier occurs at the base of the glacier or at the glacier sole.

Erosion of bedrock or pre-existing sediment at the base of the glacier occurs by 2 main processes; abrasion and plucking or quarrying. Abrasion generally involves the wearing away of rock substrate by tools, usually rock fragments (clasts), carried in the ice (Shaw 1985). These rock fragments must be brought into contact with the underlying rock for abrasion to occur. With downward pressure and dragging, the tools produce polished rock surfaces, striations (long linear scratches), and nail head and rat tail striae from which the direction of ice flow can be determined (Figure 21.6). Glacial grooves, long, wide, linear depressions, commonly containing striations, are also a product of abrasion. Subglacial meltwater may play an important role in the formation of rat tail striations and grooves as well. Where a number of point contacts are made by the tool on the bedrock surface, possibly as the result of variations in subglacial water pressures, crescentic marks or gouges form. These forms include crescentic fractures, crescentic gouges and lunate fractures, which are usually arranged in a row, or a train, parallel to the direction of ice flow.

Turbulent subglacial meltwater may explain the formation of "P" forms, or plastically sculptured forms (Dahl 1965; *also* referred to as "S" forms, J. Shaw, University of Alberta, personal communication, 1990). *Sichelwannen*, *muschelbrüken*, cavettos and comma forms, types of "P" or "S" forms (Dahl 1965; Kor et al. 1991), are commonly cut into uneven or irregular bedrock surfaces by such meltwaters.

Plucking, quarrying and large block entrainment involve the removal of blocks of material from bedrock or pre-existing sediments. The blocks usually separate along pre-existing weaknesses, such as joints and fractures. The main causes of the entrainment processes are the shear stresses generated at the base of the glacier, variations in subglacial meltwater pressures and alternating freeze-thaw-freeze plucking in leeside cavities. Landforms which are

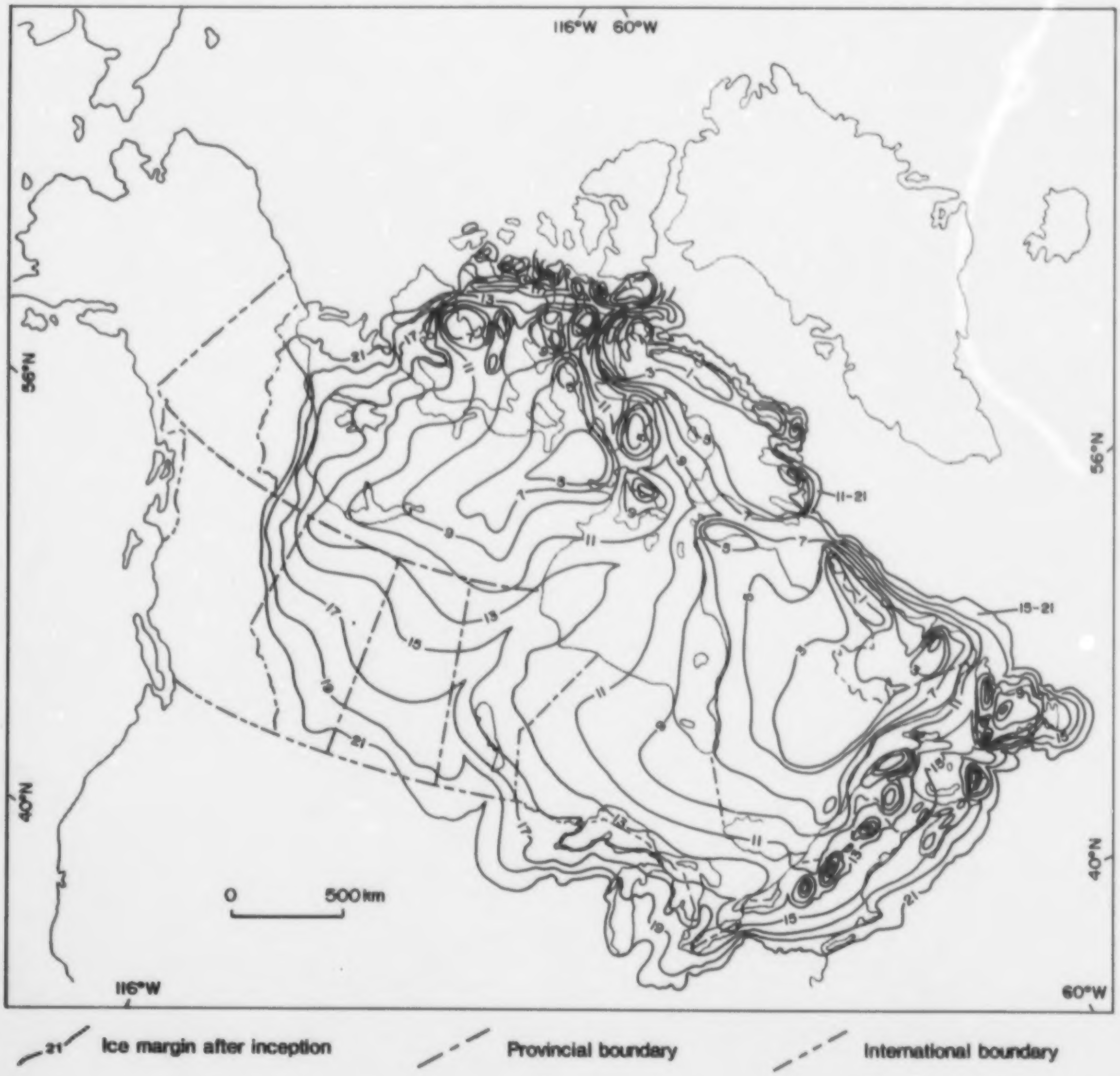


**Figure 21.5.** Hypothetical growth of the Laurentide Ice Sheet following the Sangamonian Interglacial Stage (contours in thousand of years after inception; *modified from* Vincent and Prest 1987).

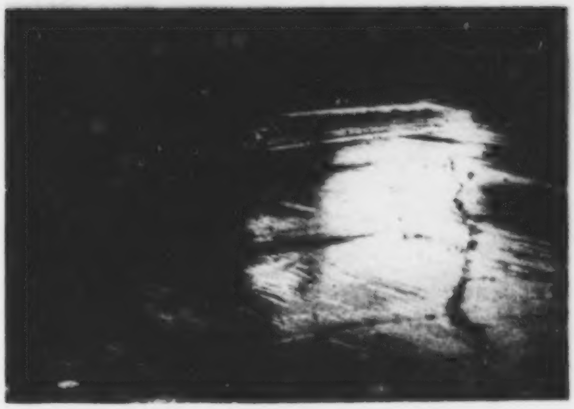


**Figure 21.6.** Striated and polished bedrock surface, Fort Frances area. Two directions of ice flow are indicated (photograph *courtesy of* A.F. Bajc).

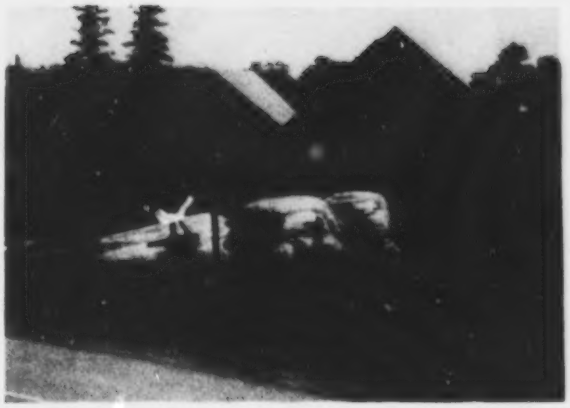


**Figure 21.7.** Small *roche moutonnée*, Renfrew County. Direction of ice flow was from right to left.

products of plucking and quarrying combined with abrasion include whale backs, stoss-and-lee forms, and *roches moutonnées* (Figure 21.7). Crag-and-tail forms and rock drumlins result from the combination of plucking, quarrying and abrasion with deposition.

Frontal incorporation, common in many polar glaciers (Shaw 1985), and supraglacial incorporation (material being incorporated from the top of the glacier), common in valley glaciers surrounded by steep mountain slopes (Sugden and John 1976), were not significant for the Laurentide Ice Sheet.

Debris, once entrained into the glacier, is transported by the glacier in 3 main zones; the supraglacial debris zone, the englacial debris zone, and the basal debris-rich zone at the very base of the glacier (Shaw 1985).

Most of the debris in continental glaciers is carried in the basal debris-rich zone or the zone of traction. This basal zone is the part of the glacier where highest strain, maximum velocity gradients and internal deformation occurs. Clast-to-clast contacts and clast-to-bed contacts are common, with abrasion being highly efficient in this zone (Shaw 1985). Clasts become faceted, striated and their edges rounded during transport within the basal debris-rich zone. Long axes of clasts are often oriented parallel to the direction of ice flow (till fabric), except in areas of highly compressive flow. Complex folding and faulting, common within this zone, help to mix and homogenize the glacial debris such that distal lithologies are mixed with more local ones (Shaw 1985). Mixing also occurs by subglacial regelation (alternating freezing and thawing) and by the mixing of sediments below the basal debris-rich zone by subglacial deformation (Boulton 1979).

Debris in the englacial zone is for the most part, highly dispersed. Clast-to-clast contact is limited as a result of low englacial strain rates and the dispersed nature and low concentration of clasts (Shaw 1985). Little modification of clasts occurs during transport. An exception to this, in continental glaciers, is where debris from the basal debris-rich zone is carried up into the englacial zone along shear planes or flow lines. Generally, the debris carried in the englacial zone has travelled farther than that in the basal debris-rich zone.

Supraglacial debris in continental glaciers accumulates primarily as a result of englacial or basal debris being transported to the surface along shear planes or flow lines in areas of compressive flow (Shaw 1985). Once at the surface, the debris is exposed to weathering and winnowing processes by which clast fracturing and sorting can occur. Fracturing produces angular clasts with little evidence of glacial transport. Constantly shifting ponds and meltwater streams on the surface of the glacier are sites of water sorting and temporary deposition of glacially transported debris (Shaw 1985). Complex mixes of sedimentological environments can result. Generally, more distant sediment and rock sources are reflected in supraglacial debris.

There are 3 main processes by which the debris carried in the glacier is deposited by the glacier; lodgement, melt-out and flowage.

The lodgement process involves deposition by pressure melting beneath active ice. This allows detrital particles to be released and lodged or plastered onto the glacier bed. This process produces tills which are gradually built up by accretion. Some type of shearing is considered essential to the lodgement process.

The melt-out process involves the slow release of debris from glacier ice that is not sliding or deforming internally (stagnant ice). This process may involve supraglacial, englacial or basal debris being gradually added to a till layer either subglacially or supraglacially by melting of the ice.

Flowage involves the downslope transport of debris-water mixtures. Sources for the mixtures include debris and meltwater previously released by the processes of lodgement or melt-out. Debris flows are most common in ice-marginal zones.

# TILL

## Introduction

Till is the name for "sediment that has been transported and deposited by or from glacier ice, with little or no sorting by water" (Dreimanis 1982a). Till is commonly poorly sorted, often containing clasts of many sizes in a variable finer-grained matrix. Till is composed of a mixture of minerals and rock types, some of which are far travelled.

Till commonly appears to be massive in structure, and to be without continuous lamination. On close examination, however, it contains many different types of structures which are clues to the processes that were active during its deposition. Till is often compact due to pressures exerted by the mass of the glacier during deposition (over consolidated). Normally consolidated or only slightly over consolidated tills can occur, particularly along the shores of the Great Lakes, where drainage of the meltwater generated during deposition was inhibited and consolidation was delayed until after the weight of the overlying ice was removed.

Till generally contains striated and faceted clasts. Elongated clasts in till may have a common orientation (till fabric) which can be related to the direction of ice movement during its deposition.

The size distribution of particles in till is largely related to the characteristics of the materials entrained by the glacier. However, a decrease in particle size does occur during transport by crushing and abrasion, primarily in the basal debris-rich zone.

Dreimanis and Vagners (1971) suggested that at least 2 modes or terminal grades exist for each rock type or lithic fragment incorporated into and subsequently crushed and abraded by the glacier; one in the clast size (lithic fragments), one in the matrix size (individual minerals). Different mineral types were found to have different terminal grades, such that a till produced by overriding granitic rocks would have a sandy textured matrix and a till produced from the crushing and abrasion of carbonate rocks would have a finer-grained silty matrix.

The composition of the clasts and matrix of till reflects the rock types that the glacier has overridden. In general, the closer the source, the higher the percentage of clasts of that source rock type. The components of the till matrix similarly reflect source parameters. This characteristic of till makes it useful as an indirect indicator of the bedrock geology in areas covered by thick till, and hence, useful in mineral exploration.

In Ontario, several components of till are studied. The lithologies of the clasts (boulders, cobbles and pebbles) are useful prospecting tools in that they reflect up-ice geology and ice-flow direction. The orientation or fabric of pebbles and cobbles is also a good indicator of ice flow direction.

The heavy mineral fraction of the till matrix is useful in determining source area and ice-flow direction. Both the mineralogy and the geochemistry of this part of the till are used in prospecting for mineral deposits.

The carbonate content and the calcite:dolomite ratio of the silt-clay fraction of tills have been found to be very useful in the characterization and correlation of tills in southern Ontario. The presence of carbonate in tills in some areas of northern Ontario has diminished the effects of acid precipitation and retarded the acidification of some of our northern lakes (Shilts 1984b).

The trace element content of the clay and the combined silt and clay fractions of till can be used in prospecting, and in the correlation and characterization of till. It also provides geochemical background values for environmental studies.

The identification of the clay mineralogy of the clay-size particles in till can be used for characterizing tills and in determining their behavioural properties. Similarly, Atterberg limits, performed on the matrix, and other physical tests are useful to engineers in determining the behavioural properties of till for construction purposes.

## Genetic Classification of Till

Studies on modern glaciers, especially those of Boulton (1968, 1970a, 1970b, 1971, 1975, 1976a, 1976b, 1978, 1979, 1980a, 1980b; Boulton et al. 1974; Boulton and Paul 1976; Boulton and Eyles 1979; Boulton and Jones 1979; Boulton and Deynoux 1981), Shaw (1976, 1980) and Lawson (1979a, 1979b, 1981a, 1981b, 1982, 1984), have led to a better understanding of the processes involved in "till" deposition. These studies have also been instrumental in identifying properties and structures in till which are diagnostic of the depositional process and can be applied to Pleistocene or older tills.

Numerous studies on Pleistocene tills have increased our understanding as well (Dreimanis 1976, 1979, 1980, 1981, 1982a, 1982b, 1983; Evenson et al. 1977; Lundquist 1977; Shaw 1977, 1979, 1982a, 1982b; Kruger 1979; Gibbard 1980; Gravenor et al. 1984; and many others). However, all of these studies have also led to confusion and a profusion of terms, until the basic definition of till itself is in question (Lawson 1979a, 1979b; Boulton 1980a; Dreimanis 1982a).

The definition proposed by the International Union for Quaternary Research (INQUA) Commission Work Group (1) will be followed in this chapter: "Till is a sediment that has been transported and deposited by or from glacier ice, with little or no sorting by water" (Dreimanis 1982a).

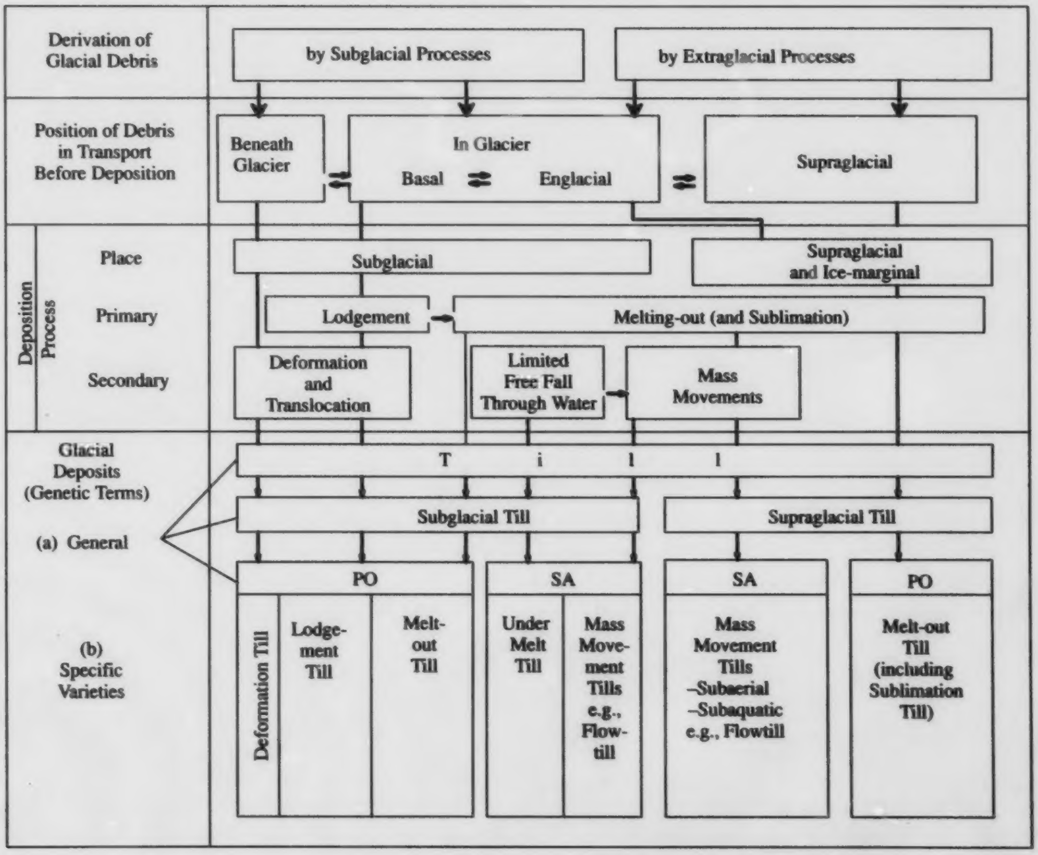
This commission produced several comprehensive and useful tables for the classification of till and how to recognize each different till type (Dreimanis 1988).

In general terms, however, till can be divided into 2 main types (Figure 21.8): 1) subglacial tills, deposited by lodgement and melt-out processes at the base of the glacier with properties related to those inherited during transport in the basal debris-rich zone; and 2) supraglacial tills, deposited by flow from the upper surface of glaciers with properties developed during transport in the englacial and supraglacial zones. It should be noted that debris in the glacier can move between the various transport zones. This twofold division is probably sufficient for drift prospecting where it is more important to recognize that the till is derived from material transported in the basal debris-rich zone, rather than identifying the processes of deposition. However, for geologists, engineers and planners it may be necessary to recognize or to understand the significance of the various genetic types of till.

Lodgement till is a subglacial till deposited by the lodgement process. It involves debris that has been transported in the basal debris-rich zone. During lodgement, clasts drag across the substrate creating striations or gouges in rock; in softer substrate the clasts plough deep grooves or flutes. Clasts are imbricate, dipping up-glacier with long axes parallel to flow; generally, this till type has a strong fabric. The bottom of the clasts become striated as they are being dragged along by the ice prior to deposition. Once lodged, clasts are abraded and striated on their tops during the time they remain at the glacier-bed interface. Bullet-shaped boulders, typical of lodgement till, are produced by this abrasion occurring after deposition. Shear planes may be preserved in lodgement till and fissility may be developed (Figure 21.9). The lower contact of a lodgement till is sharp and erosional, and often truncates bedding in underlying sediments.

Sand lenses and stringers can be quite common in lodgement till. They form during the periodic draining of subglacial meltwater in thin films or within cavities beneath the ice. Subglacial drainage of meltwater can also occur by a network of Nye- or n-channels. Once infilled, these become channel-shaped ribbons of sand and gravel within the till. These channels have sharp lower contacts and erosional upper contacts that are often truncated or sheared.

In the subglacial environment, till deposition also takes place by melt-out. Preservation of englacial structures may occur as the result of the slow release of debris from stagnant glacier ice. Most subglacial melt-out tills originate from the basal debris-rich zone of glaciers; therefore, clasts are striated and faceted. Many of the grain size characteristics of the debris in transport are preserved. When compared within the same area, melt-out tills are coarser grained than lodgement tills because abrasion continues to occur during the lodgement process. Clasts of subglacial meltout till are usually oriented parallel to ice flow as in lodgement tills;



Derivation of Glacial Debris
by Subglacial Processes
by Extraglacial Processes
Position of Debris in Transport Before Deposition
Beneath Glacier
In Glacier
Basal
Englacial
Supraglacial
Deposition Process
Place
Primary
Secondary
Subglacial
Supraglacial and Ice-marginal
Lodgement
Melting-out (and Sublimation)
Deformation and Translocation
Limited Free Fall Through Water
Mass Movements
Glacial Deposits (Genetic Terms)
(a) General
(b) Specific Varieties
T i l l
Subglacial Till
Supraglacial Till
PO
SA
SA
PO
Deformation Till
Lodgement Till
Melt-out Till
Under Melt Till
Mass Movement Tills e.g., Flow-till
Mass Movement Tills –Subaerial –Subaquatic e.g., Flowtill
Melt-out Till (including Sublimation Till)

**Figure 21.8.** Genetic classification of till. Factors that influence the formation of till shown in the upper half, and a tentative depositional genetic classification of tills shown in the lower half (*modified from* Dreimanis 1982a). PO - primary or ortho-tills; SA - secondary or allo-tills.

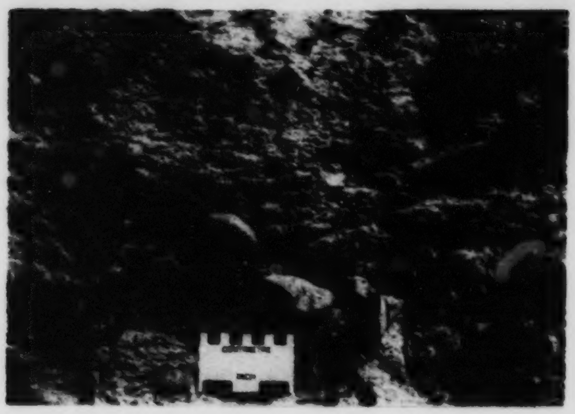


**Figure 21.9.** Lodgement till containing shear planes (s) and a weakly developed fissility (f), Coldwater.

however, clasts in melt-out tills usually dip at lower angles up-glacier than do clasts in lodgement tills.

Large volumes of water must be evacuated through the till during the formation of subglacial melt-out tills. Silt or sand coatings around clasts and thin irregular layers of sorted sediment may form along drainage passageways. Loss of volume on melting leads to draping of sediment layers over clasts (Figure 21.10). Boulder scours are associated with melt-out till.

Supraglacial melt-out tills are rare in the Pleistocene record because once the debris is released at the surface of the glacier, it is highly susceptible to being mobilized and deposited as flowtill.

Debris flows or sediment gravity flows are a product of the downslope movement of debris. If these flows can be associated with a glacier, they are referred to by some authors as flowtills (Dreimanis 1982a). Flowtills can be massive and poorly sorted. More commonly, they are associated with stratified sediments (Figure 21.11). Clast fabrics are not as strong as in the other 2 types of till; when compared regionally, clast fabrics in flowtills are quite variable. Flowtills may contain: floating or rafted boulders, flow noses, rip-up clasts, flow shears, and load and dewatering structures (Dreimanis 1988).

Flowtills can be of subglacial or supraglacial origin; the latter type is more common. Flowtills are commonly associated with ice-marginal positions and occur frequently in moraines.

Deformation till forms as the result of extensive subglacial deformation. The resultant till is commonly massive and homogeneous when fine grained (Figure 21.12). It may be the dominant type of till in glaciated basins, where the fine-grained nature of the till (due to the incorporation of fine-grained glaciolacustrine sediments) and moisture content lead to low shear strength, possible failure or deformation (Barnett 1987). In coarse-grained sediments, deformation does not appear to be as pervasive.

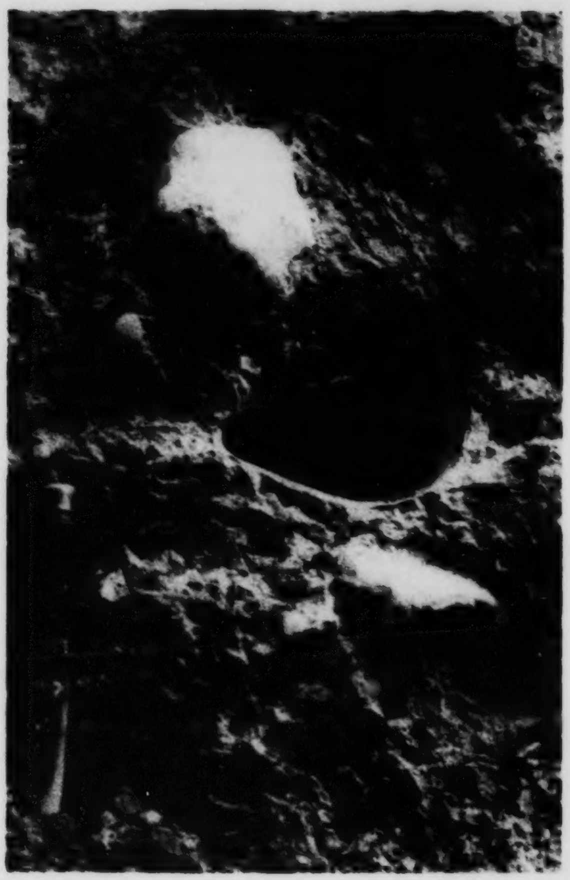


**Figure 21.10.** Subglacial melt-out till with silt and sand coatings and lenses around clasts, near St. Thomas. (Knife is about 20 cm long.)

## Landforms Associated with Till

The landforms created by a continental ice sheet associated with till can be divided into: 1) linear features parallel to ice flow, 2) linear features transverse to ice flow, and 3) features lacking consistent orientation (Sugden and John 1976; Prest 1983).

Linear features parallel to ice flow include streamlined subglacial forms, such as fluted or drumlinized ground moraine or till plains, drumlins or drumlinoid ridges and crag-and-tail features. Linear ice-marginal forms oriented parallel to the direction of ice flow include some interlobate moraines.

There are many suggested mechanisms for the formation of fluted or drumlinized ground moraine (Figure 21.13). The most common is that the small flutes form by the squeezing of saturated debris into linear cavities that formed in the lee of large boulders during lodgement (Dyson 1952). Alternatively, flutes are formed by the migration of debris beneath the glacier as a result of differential pressure gradients (Galloway 1956; Boulton 1971).

Drumlins, oval hills composed of till or stratified sediments, are also controversial in their genesis. Genetic

**Figure 21.11.** Small nose of a flow composed predominantly of flowtill near Plum Point, about 50 km southwest of London (photograph *courtesy of* P.J. Barnett *from* Sharpe and Barnett 1985).

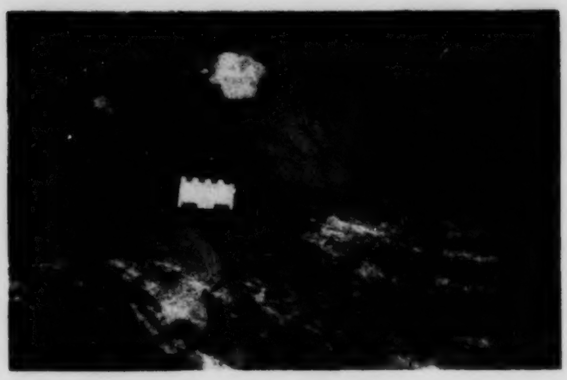


**Figure 21.12.** Deformation till exposed along the Lake Erie shore bluffs near Port Burwell. Glaciotectonic deformation involves both the underlying silts and clays and the lower part of the deformation till.

hypotheses for drumlins include: 1) development through changes in the geotechnical properties of subglacial debris as a result of pressure variations (Smalley and Unwin 1968), 2) accretion of till around an obstacle such as a bedrock protrusion (Boulton 1970b, 1971), 3) formation as a result of differential pressures at the base of the glacier (Evenson 1971), or 4) helical motion of flow within the ice (Shaw and Freshauf 1973). A cavity-filling hypothesis (Shaw 1983) suggests that large subglacial meltwater discharges can carve cavities into the base of the ice which are similar to flute marks, but the inverse; these cavities are then filled with stratified sediment during the waning stages of flow. Shaw and Sharpe (1987) have suggested that major subglacial discharge events can cut drumlin forms out of pre-existing drift.

Some interlobate moraines form along an initial zone of weakness in the glacier, prior to the actual break up and formation of separate lobes, and therefore can be oriented parallel to the iceflow direction. More commonly, however, interlobate moraines form perpendicular to the direction of ice flow, between adjacent ice lobes or during fluctuations when the ice lobes resume contact. Interlobate moraines are usually composed mainly of stratified sediments. Flowtills are common. In addition, because of the nature of these landforms, several subglacial till layers may interdigitate with the stratified sediment core.

Linear features that form transverse to ice flow include the subglacially formed Rogen or ribbed moraine and De Geer or washboard moraine. Rogen moraines are fairly large-scale, transverse lineaments which give an overall ribbed appearance to the land surface. Cowan (1968) suggested Rogen moraines formed by pushing or overriding of existing debris by reactivated ice. Lundqvist (1969) suggested Rogen moraines formed by overriding crevasse fillings formed in a zone of tension beneath the ice. Compressive flow mechanisms have also been proposed. Bouchard (1988) suggested Rogen moraines formed as a result of the shearing and stacking of near-base englacial debris which later becomes deposited as basal melt-out till.

De Geer moraines are ridges of till and stratified sediments that probably form at the grounding line of an ice shelf when the glacier margin is in a large, deep body of water. De Geer moraines consist of a succession or series of regularly spaced, discrete, narrow ridges seldomly exceeding 15 m in height and usually spaced less than 300 m apart (Sugden and John 1976). Individual ridges of De Geer moraine may have formed annually; however, there remains much debate on this suggestion (*see* Sugden and John 1976). De Geer moraines are well developed in northwestern Ontario, particularly in the glacial Lake Agassiz basin.

Ice-marginal forms include end moraines, push moraines and ice-thrust moraines. The sediments found in these features are quite variable, including all genetic varieties of till, stratified sediments and deformed pre-existing sediment and rock. End moraines are the common form in Ontario, with over 45 identified (Taylor 1913; Chapman and Putnam 1984; Zoltai 1961, 1965; Boissonneau 1966, 1967). Some of the larger end moraines in southern Ontario are up to 190 km long (the Wyoming moraine of the Huron lobe), several kilometres wide and stand up to 30 m above the surrounding landscape (Karrow 1989). The Lac Seul moraine in northwestern Ontario exceeds 600 km in length (Zoltai 1965). Other end moraines are considerably smaller, in the order of a few kilometres long; some have little or no topographic expression (Barnett 1985; Karrow 1989).

Ground moraine and hummocky ground moraine are 2 subglacially formed moraine types lacking consistent orientation. Local relief on both of the moraine types is low, between 2 and 10 m (Prest 1983). Crevasse fillings and ice-pressed forms can occur as random or rectilinear ridges. Crevasse fillings can exceed 7 m in height and 0.5 km in length (Gravenor 1957). Crevasse fillings have been described in the Barrie area (Deane 1950), the Lindsay–Peterborough area (Gravenor 1957) and associated with the St. Thomas moraine near Port Stanley (Dreimanis 1972). Crevasse fillings are composed of a variety of sediments including till, glaciofluvial sand and gravel, and glaciolacustrine sand, silt and clay (Deane 1950; Dreimanis 1972). High-relief hummocky moraine, which is formed as the glacier surface downwastes rather than retreats, is generally referred to as disintegration moraine (Gravenor and Kupsch 1959; Sugden and John 1976). The Dummer moraine, located along the southern shores of the Kawartha Lakes, is an extensive area of hummocky topography composed of bouldery till, sand and gravel. It may be the only example of disintegration moraine of any extent, in Ontario.

Landforms in southern Ontario are discussed in *The Physiography of Southern Ontario* by Chapman and Putnam (1984); major landforms of northern Ontario are discussed in papers by Zoltai (1965), Boissoneau (1966, 1967), Shilts et al. (1987) and Dredge and Cowan (1989).

## Glaciofluvial, Glaciolacustrine and Glaciomarine Deposits and Landforms

### INTRODUCTION

Considerable volumes of meltwater are generated by and discharged from glaciers. Meltwater production at modern glaciers has been found to be cyclical: a yearly cycle that is associated with the summer release of water stored in the glacier during the fall, winter and spring months; and, a diurnal summer cycle which is associated with warmer daytime temperatures and cooler temperatures at night. The

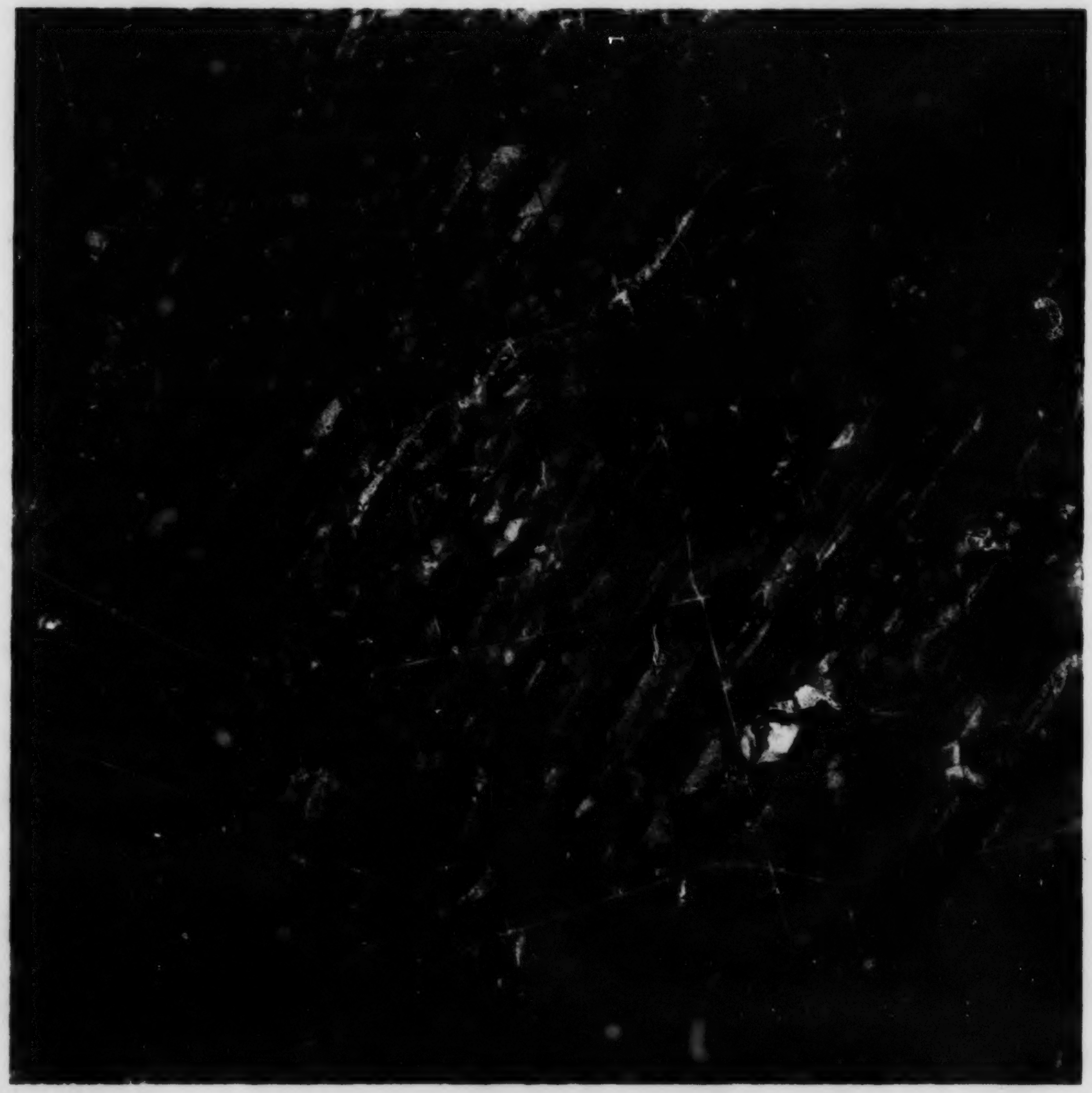


**Figure 21.13.** Drumlinized till plain near Peterborough (National Air Photo Library A24314-29).

diurnal fluctuations, however, were probably more subdued in the large Pleistocene ice sheets (Smith 1985).

Superimposed on these cycles of meltwater discharge are the effects of rainfall or long periods of sunshine, which periodically increase the discharge of meltwater from the glacier. A rare fluctuation in meltwater discharge may be caused by the sudden release of large amounts of meltwater stored in englacial or subglacial cavities, subglacial depressions or in ice-dammed marginal lakes. Sometimes catastrophic, these outbursts of meltwater, called "jökulhlaups", are recorded by a gradual rise in discharge which peaks sharply and is followed by a rapid decline over a period of hours or days (Smith 1985).

Meltwater within the ice is mainly drained by a network of englacial and subglacial tunnels or conduits. Subglacial conduits which cut upward into the ice are called "Röthlisburger channels" and those which cut into the underlying substrate are termed "Nye-channels". The network of tunnels that develop within glaciers are probably not capable of handling abrupt increases in flow, such as those that occur during jökulhlaups (Shreve 1972). Sheet flow along the base of the glacier probably occurs during jökulhlaups.

Meltwater flows away from the front of the glacier in rivers and streams, where the topography permits. These meltwater streams eventually flow into lakes (non-contact glacier-fed lakes) or into the sea. Where the drainage of meltwater away from the ice is hindered by topography or sea level, meltwater from the glacier can issue directly into lakes (ice-contact glacier-fed lakes) or into the sea.

Large amounts of glacial debris are transported by meltwater and deposited under the glacier in conduits and cavities and beyond the ice margin in rivers, streams, lakes and seas. The glacial debris is no longer till-like for it becomes disaggregated, better sorted, and stratified by the meltwater currents during transport and deposition.

Several erosional features and landforms have been related to subglacial meltwater. These include "P" or "S" forms, potholes, drumlins and tunnel valleys (Sugden and John 1976; Shaw 1985). The formation of "P" or "S" forms and drumlins has been related to the action of turbulent subglacial sheetflows that formed during catastrophic releases of subglacially stored meltwater (Shaw 1983, 1985; Shaw and Sharpe 1987; Kor et al. 1990, 1991). Tunnel valleys, which are also formed as a result of erosion by catastrophic discharges of meltwater beneath glaciers, can be several kilometres wide and hundreds of metres deep (Sugden and John 1976; Barnett 1990). Steep, streamlined valley walls and the occurrence of eskers and drumlins within the tunnel valleys are criteria used to suggest a subglacial origin for these valleys.

## GLACIOFLUVIAL DEPOSITS AND LANDFORMS

Deposits from meltwater in rivers or streams that flow on, within, beneath or beyond the glacier are referred to as glaciofluvial deposits. Glaciofluvial deposits are further subdivided into ice-contact stratified drift and outwash.

Ice-contact stratified drift is deposited beneath, on, within or immediately adjacent to, the glacier where the influence of the glacier remains strong. Outwash is deposited beyond the ice margin in the rivers and streams issuing from the glacier. Direct glacier influence is not as strong. These environments of deposition produced deposits and landforms with distinct characteristics. This subdivision of glaciofluvial deposits is quite practical, in particular with regard to their predictability as mineral aggregate resources.

### Ice-Contact Stratified Drift and Associated Landforms

Ice-contact stratified drift is the general term given to sediments deposited by glacial meltwater in contact with the glacier. Ice-contact stratified drift deposits are often quite variable both laterally and vertically (Figure 21.14). This is the result of the highly variable distribution of meltwater beneath glaciers, the extreme variability in the rate of discharge of meltwater beneath glaciers and the tendency for subglacial drainage to be rapidly shifted from one tunnel system to another.

Predominantly composed of discontinuous layers of sand and gravel, ice-contact sediments can contain silt and clay, and till or flowtill layers. Secondary structures such as faulting and folding are common, often resulting from the removal of ice support. Deposition occurs within cavities, crevasses and tunnels within or under the ice, within ice-walled channels on the surface of the ice, between 2 ice lobes or along the glacier margin.

Commonly, landforms composed of ice-contact stratified drift are positive relief forms and include eskers, kames, kame terraces, crevasse fillings, interlobate moraines, ice-marginal deltas, subaqueous fans and some end moraines and drumlins (Sugden and John 1976).

Eskers are casts of tunnels carved into the glacier by meltwater. They are usually long sinuous ridges that roughly parallel the direction of local glacial retreat (Figure 21.15). Eskers commonly consist of a core of gravel and sand and are flanked by sands. Cyclic sedimentation patterns have been recognized within esker core sediments, which relate to variations in meltwater discharge through the conduit (Shaw 1985). Faulting occurs in sediments deposited along the margins of eskers that formed in subglacial tunnels upon the removal of the supporting walls of ice during melting. Extensive faulting in an esker may suggest that it had formed in an englacial tunnel or in an ice-walled channel on the glacier surface. This faulting develops during the subsidence of the ice-contact sediments to the ground surface during melting of the supporting ice.

Kames are isolated hills composed of ice-contact stratified drift. Having been formed in contact with the ice, collapse structures such as faults, folds and steeply dipping beds are common. Kames are usually small and have a tendency to be predominantly composed of sand.

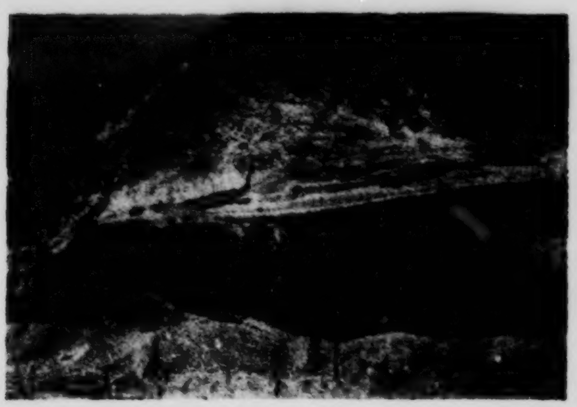


**Figure 21.14.** Ice-contact stratified drift found within an esker located in Renfrew County. Note the variation in grain size and the discontinuous bedding.

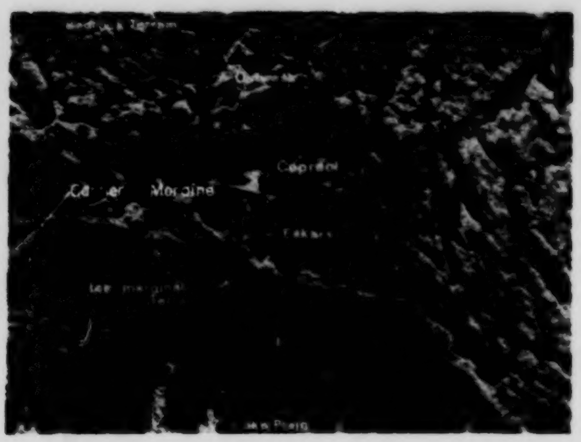


**Figure 21.16.** Radar image of an ice-marginal delta formed along the Cartier I moraine at Capreol (image by RADARSAT, Canada Centre for Remote Sensing).

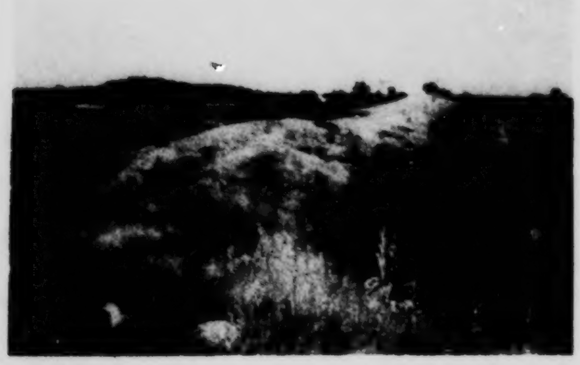


**Figure 21.15.** The Tweed esker, about 10 km south of Tweed.

Kame terraces or ice-contact terraces are formed between the glacier and a confining topographic high. The sediments are similar to those found in outwash deposits, except that they can be extensively disturbed along the ice-contact side.

Ice-marginal deltas and subaqueous fans form where the conduits issue into standing bodies of water, such as ice-contact glacier-fed lakes and seas (Figure 21.16). These deposits are included in ice-contact stratified drift because a number of their properties are greatly influenced by the ice-marginal and conduit environment. Ice-marginal deltas commonly have an up-glacier ice-contact slope. Their internal composition is similar to Gilbert-type deltas with topset, steeply dipping foreset and bottomset beds. Frontal profiles are usually steep and sediment grain size and sedimentation rates decrease lakeward (Bogen 1983).

Subaqueous fans form where the conduit mouth is located below water level. Subaqueous fan deposits of gravel and sand tend to fine radially outward from the conduit mouth and the final deposit often becomes finer grained upward as well (Rust and Romanelli 1975; Cheel and Rust 1982; Shaw 1985).

Interlobate moraines, sometimes referred to as kame moraines, are formed between 2 ice margins or within a re-entrant in the glacier. Interlobate moraines commonly consist of a series of coalescing subaqueous or subaerial fans or kames that are ice supported on both sides. Fine sands and silts of glaciolacustrine origin are common in interlobate moraines such as the Oak Ridges and Orangeville moraines (Cowan 1976; Duckworth 1979). Interlobate moraines, being deposited at the margin of 2 ice masses, may have complex and unpredictable stratigraphy. Crevasse fillings, drumlins and some end moraines may contain areas of ice-contact stratified drift.

## Outwash Deposits and Associated Landforms

Outwash is the general term for glaciofluvial sediments deposited in rivers and streams beyond the glacier margin. Meltwater streams that issue from glaciers have very high sediment loads as a result of the large amount of readily available, loosely consolidated glacial debris under, at or near the ice front. Most of this debris is transported in shallow, braided stream channels, which are continually shifting across a sometimes broad alluvial plain in response to the highly variable flow rates typical of meltwater (Smith 1985). A topographically constricted alluvial plain is called a "valley train" or a "valley sandur". An unconfined alluvial plain is called an "outwash plain" or a "plain sandur". Characteristics of outwash deposits formed in either setting are fairly similar.

Outwash deposits are coarse grained, proximal to the ice margin, and generally decrease in grain size downstream (Figure 21.17). Sedimentation occurs in various types of unit and braid bars in the proximal zone, and produces gravel deposits that are predominantly massive to crudely horizontally stratified (Smith 1985). Preservation of finer grained-sediments deposited during waning flows is rare. If preserved, the fine-grained sediments are usually found in thin, discontinuous beds.

Farther downstream, grain size decreases and more tranverse-type bars are present. Tabular sets of planar cross-beds become more abundant along with beds of cross-laminated and horizontally bedded sands (Figure 21.18). Grain size changes more gradually in outwash deposits than in deposits of ice-contact stratified drift and is generally more laterally consistent.

Kettle holes and kettle lakes may form in the proximal zone of the sandur where parts of the glacier become buried beneath the outwash sediments and later melt. Also, large blocks of ice can become stranded on the sandur surface during jökulhlaups. These later melt and form kettle holes and lakes. The terms "kettled sandar" or "pitted outwash plains" are used for these landforms.

Where the meltwater streams enter lakes, deltas form in response to flow expansion and the loss of flow velocity and transporting capacity (competence). The deltas are commonly Gilbert-type deltas with well-developed topset, foreset and bottomset beds. Deltas fed by meltwater streams are included in outwash deposits on the accompanying maps (*see* Maps 2553, 2554, 2555 and 2556, map case). The strong influence of fluvial processes on these deposits, combined with the similarities in grain size and primary sedimentary structures of delta topset beds and outwash deposits, favours the inclusion of deltas in glaciofluvial outwash deposits.

## GLACIOLACUSTRINE DEPOSITS AND ASSOCIATED LANDFORMS

Sediments that have been carried by glacier meltwater and subsequently deposited in lakes are referred to as glaciolacustrine deposits. The lake may be in contact with the glacier (an ice-contact glacier-fed lake) or may be fed by glacier meltwater streams (non-contact or distal glacier-fed lake; Smith and Ashley 1985). Both types of glacier-fed lakes are affected by the characteristics of glacier meltwater, including: the strong seasonal- and weather-dependent discharge, the low water temperatures, and in particular, the high sediment loads (Smith and Ashley 1985).

Where meltwater streams or rivers enter lakes, sudden flow expansion causes an abrupt decrease in stream velocity and competence. Rapid deposition of the streams' coarser load results and a Gilbert-type delta forms. Sediment is distributed along the delta front by avalanching, suspension fallout, underflowing currents and gravity-induced deformation (Smith and Ashley 1985). The finer-suspended sediment is carried farther out into the lake basin. The lake basin sediments generally become gradually finer grained away from the delta; sand is deposited near the sediment input source, through silt-dominated sediments to clay-dominated sediments.

The way in which the inflowing glacier meltwater interacts with the lakewater affects the transport and deposition of the suspended sediment load and the finer-textured bed load. Density differences between the inflowing water and lakewater and the density profile in the lake itself determine if the sediment plume will become an underflow, interflow or overflow (Figure 21.19). The high sediment loads of glacial meltwater streams commonly produce underflows. Interflows can occur, however, in strongly density-stratified glacier-fed lakes which are not in direct contact with the ice.

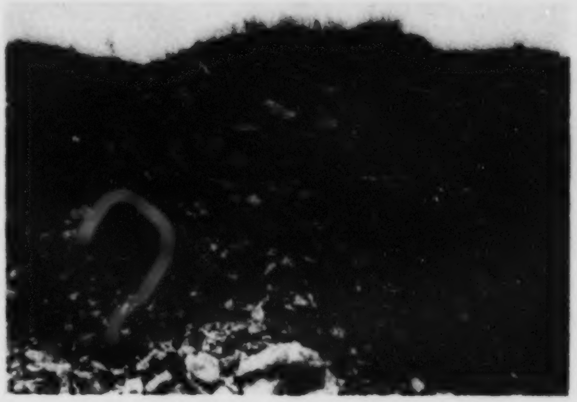


**Figure 21.17.** Proximal outwash gravel, Eganville.

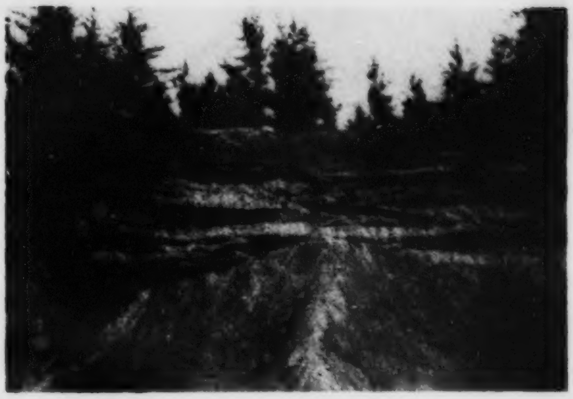


**Figure 21.18.** Cross-bedded outwash sand and gravel deposits, Kirkland Lake area (photograph *courtesy of* C.L. Baker).

The most common types of glaciolacustrine sediments deposited within glacier-fed lake basins are rhythmites. Rhythmites consist of fining-upward sedimentation units. Each unit consists of a couplet composed of a sand-silt base overlain by a silt-clay layer. Rhythmites produced in density-stratified glacier-fed lakes show proximal to distal thinning and fining as a result of sediment dispersal by interflows (Smith and Ashley 1985). These rhythmites are distributed at nearly all elevations within the lake basin. Rhythmites produced solely by underflows, however, are distributed only in the lower parts of the lake basins and are absent from topographic highs. The thicknesses of the summer sand-silt layers are also controlled by lake bottom topography and sediment input volume.

The coarse, lower part of the rhythmite can be the dominant component in the proximal zone, for example, in the lower parts of large sand-dominated, low frontal-profile deltas, the fine sand or silt rhythmites are commonly several metres thick (Figure 21.20; Ashley 1975; Barnett 1987). Farther out in the lake basin, the clay or fine part of the rhythmites is usually thicker than the coarse part.

The overall cyclicity in lake sediments is the result of the input of seasonal variation in sediment supply. Varves are rhythmically bedded sediments which record this yearly cycle (De Geer 1912). Rhythmites may also be produced by a single underflow, slump-generated surge currents, diurnal discharge variations and weather-related discharge variations (Smith and Ashley 1985). These types of rhythmites are not varves.

In ice-contact glacier-fed lakes, the proximity of the glacier greatly affects the depositional environments. The glacier disrupts the thermal-density stratification in the lakes, preventing the development of overflows and interflows. The input of meltwater and sediments from conduits in the ice below water level creates subaqueous fans instead of deltas. The direct input of glacial debris into the lakes produces flowtills or debris flows. A calving ice margin produces icebergs, which may scour lake bottoms and disturb existing bottom sediments. The icebergs can also distribute glacial debris throughout the lake as dropstones and debris dumps.

Along the margins of lakes, wind-generated waves and currents modify the sediments exposed along the shore and in the nearshore area. Beaches, spits, bars and shore bluffs are created. The beach bars and spits are commonly composed of sand and gravel; offshore bars are commonly composed of sand. The composition and texture of shore and nearshore sediments is highly dependent on the type and composition of the sediments being eroded in the shore bluffs and in the nearshore zone, the strength of waves and currents acting on the shore, and the grain-size distribution of sediments entering the lake by way of rivers and streams.

Today, abandoned deltas, wave-cut shore bluffs (Figure 21.21), beaches, spits, bars and lake plains mark the former locations of glacial lakes. The abandoned shoreline features are now tilted, higher in elevation at the northern end of an ancestral lake than at the southern end (Figure 21.22); the result of differential isostatic rebound or recovery from the mass of the continental glaciers.

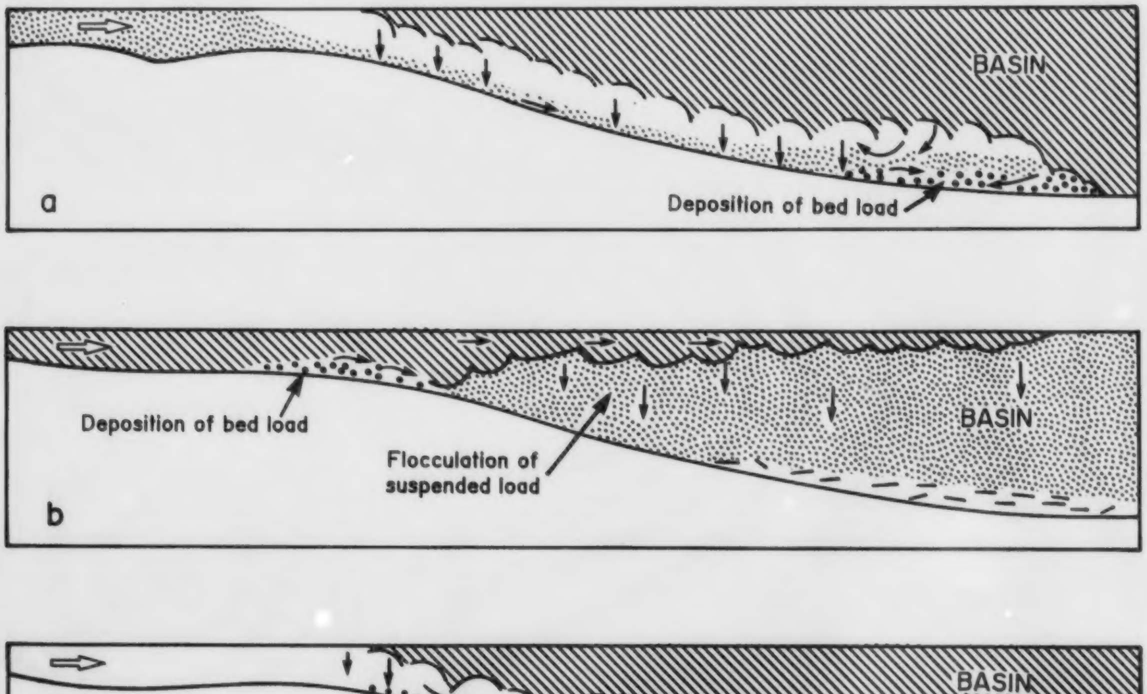


**Figure 21.19.** Modes of interaction between sediment-laden river waters and lake basin waters, determined by the relative density of the water bodies: a) underflow currents; b) overflow currents; and c) interflow currents (*modified from* Elliot 1980).

Glaciolacustrine clay deposits have been used as raw material for the production of drainage tile and brick (Guillet 1977). Today, brick is produced primarily from shales in Ontario; tiles produced from clay are slowly being replaced by plastic tiles.

## GLACIOMARINE DEPOSITS AND ASSOCIATED LANDFORMS

Glaciomarine deposits and landforms are similar to those of glaciolacustrine origin. The main differences arise from the salt in seawater. The salt in seawater makes it denser than lakewater. The greater density may result in overflow and interflow currents being more common in seawater than in lakewater. In addition, the salt content of seawater causes fine particles in suspension to flocculate, aggregate and settle more readily than in the glaciolacustrine environment. This process produces deposits of massive, silty clay and clay rather than the rhythmites of the glaciolacustrine environment. Rhythmites do form in the glaciomarine environment; however, they appear to be confined to lower delta foreset and proximal bottomset deposits, and to the distal zone of subaqueous fan deposits.

Fine-grained marine and glaciomarine sediments may develop a metastable fabric through the flocculation stage of the depositional process. Glaciomarine and marine deposits that have this metastable fabric are referred to as "quick" or "sensitive" clays. These types of sediments are prone to failure when the metastable structure collapses as a result of saturation and physical disturbance.

## ONTARIO'S RECORD OF GLACIATION

### Introduction

The deposits and effects of glaciation are widespread throughout Ontario. The rounded, smoothed and abraded surfaces of rock knobs of the Canadian Shield, the large decorative pink and grey boulders in our parks and playgrounds, the hummocky ridges of unconsolidated sediments, and the expansive level regions of clay and sand are all evidence of the former presence of glaciers in Ontario. It is from similar pieces of evidence that the theory of widespread continental-scale glaciation itself was founded less than 200 years ago (*see* Imbrie and Imbrie (1979) for an excellent account of the development of glacial theory).

The general distribution of the various types of Quaternary sediments and the major landforms associated with them are displayed on the accompanying maps (*see* Maps 2553, 2554, 2555 and 2556, map case). These maps are compiled from Quaternary, engineering terrain and surficial geology maps. The level of detail, and the reliability of the maps in general, increases towards the south where the deposits of glaciation are thick and extensive, and where development pressures on these resources were first observed.

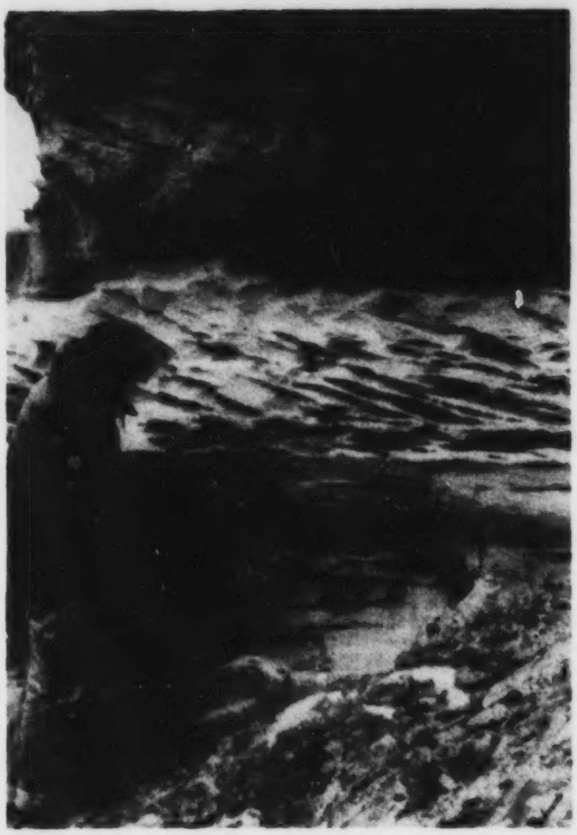


**Figure 21.20.** Sand and silt rhythmites, west of Port Burwell. Coarse part of the rhythmite (a) is very fine sand with climbing ripples; the fine part (b) is clay.

**Figure 21.21.** Wave-cut shore bluff, near Port Elgin.

This section discusses the characteristics of the various geological units shown on the accompanying Quaternary geology maps (*see* Maps 2553, 2554, 2555 and 2556, map case). The section also describes significant Quaternary deposits which do not outcrop or are too limited in extent to be shown at the map scale of 1:1 000 000. For detailed

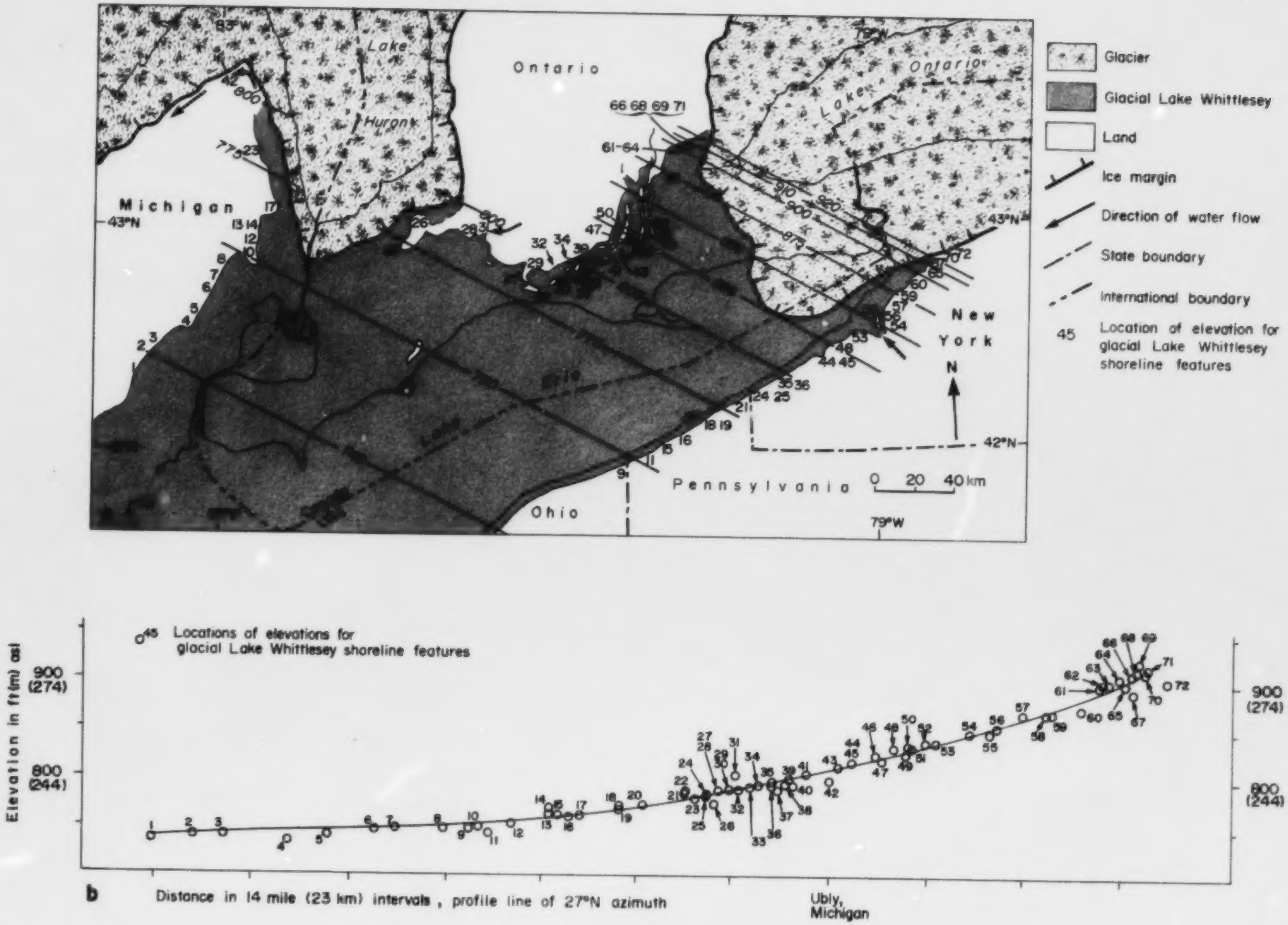

Ontario
Michigan
43°N
83°W
Lake Huron
Lake Ontario
79°W
43°N
New York
N
42°N
Pennsylvania
Ohio
0 20 40 km
79°W
Glacier
Glacial Lake Whittlesey
Land
Ice margin
Direction of water flow
State boundary
International boundary
45 Location of elevation for glacial Lake Whittlesey shoreline features
66 68 69 71
61–64
Elevation in ft(m) asl
900 (274)
800 (244)
45 Locations of elevations for glacial Lake Whittlesey shoreline features
b
Distance in 14 mile (23 km) intervals, profile line of 27°N azimuth
Ubly, Michigan

Figure 21.22. Tilted shoreline of glacial Lake Whittlesey: a) plan view; b) cross section drawn parallel to the direction of maximum tilt of the shoreline (*modified from* Barnett 1979).

information on specific deposits or areas, please refer to the larger scale maps and reports upon which this compilation is based (*see* Figure 1 *in* Sado and Carswell 1987; OGS 1990a, 1990b and references cited within the text).

## Geological Setting

The sediments deposited in Ontario during the Quaternary lie on much older rocks of Precambrian, Paleozoic and Mesozoic age which have been discussed in previous chapters. Bedrock geology and structure played major roles in the development of the landscape of Ontario.

Ontario can be divided into 3 main physiographic regions, based mainly on variations in bedrock geology. A central area of uplands, underlain by Precambrian granites, gneisses and metavolcanic and metasedimentary rocks of the Canadian Shield, separates 2 major lowland areas. The upland area, with its many ridges and valleys, formed as a result of bedrock structure and differential erosion.

The lowland to the north of the upland area, the Hudson Bay Lowland, is underlain by relatively flat-lying sedimentary rocks of Paleozoic and Mesozoic age. The lowlands to the south, the Great Lakes–St. Lawrence Lowlands, are underlain by gently dipping Paleozoic sedimentary rocks.

Bostock (1970a, 1970b) has further subdivided these 3 physiographic regions based on elevation, relief and rock type (Figure 21.23). Chapman and Putnam (1984) further subdivided the Great Lakes–St. Lawrence Lowlands and the southern part of the Precambrian Uplands into several physiographic regions based on the additional criteria of surficial material type and glacial landform.

Prior to the Quaternary Period, the bedrock of Ontario was subjected to a long period of subaerial exposure and weathering. Extensive river systems developed on the preglacial land surface with the main or master valleys being located in the outcrop belts of rock formations that were the least resistant to erosion (Fenneman 1938; Hough 1958). The location of only fragments of these preglacial river systems are known today. Evidence of their existence was either destroyed during glaciation or was deeply buried by glacial deposits. The best known buried valley system is that of the Laurentian River, which has been traced between Georgian Bay and Lake Ontario (Spencer 1907; White and Karrow 1971; Eyles et al. 1985). Parts of several other buried valleys are also known (Karrow 1989).

The present morphology of the bedrock surface has not changed substantially from that which existed before glaciation. Much of the Canadian Shield topography developed in pre–Paleozoic time (Ambrose 1964) and has survived glaciation with only minor alterations. The major landform features of the sedimentary rocks, such as the Black River, Niagara, Onondaga and Ipperwash escarpments, also have preglacial origins. Generally, it is believed that only a few metres to a few tens of metres have probably been removed from the bedrock surface by glacial erosion (Shilts et al. 1987; Dredge and Cowan 1989).

Glacial erosion may have, however, deepened and broadened preglacial valleys that were developed on wide outcrop belts of soft, erosion-prone rock types. This probably occurred in the lower Great Lakes area, where Ordovician and Devonian shales underlie the basins of Georgian Bay and lakes Ontario and Erie.

An important aspect of the preglacial landscape is its topography. Chamberlin (1883, 1888) first recognized the influence of landscape topography on the shape of the ice margin and the direction of ice flow. Chamberlin suggested that "a broad depression, reaching well backward along the line of glacial movement, was effective in producing prolongations of the ice [ice lobes], while narrow valleys, even if deep, were comparatively ineffectual, if they were associated with rough topography or if their courses were tortuous, or transverse, or oblique to the general ice movement" (Chamberlin 1888, p.183–184).

Lobation of the ice margin occurred in response to the broad basins of the Great Lakes. This produced suites of tills with distinct properties in each of the Great Lakes basins, reflecting the properties of the rock types within and up-glacier from the basin. In the interbasin areas, such as the central upland of southwestern Ontario known as the "Ontario Island" (Taylor 1913), complex stratigraphies developed as the result of oscillating ice lobe margins in an interlobate zone.

The location of drainage divides in relation to the ice sheet margins is also very important; conditions at the base of the ice sheet, the processes of deposition and hence the type and characteristics of the sediments deposited can all be affected. In general, where a glacier margin is advancing up-slope or into a basin, there is a possibility for ponding of meltwater in front of the glacier and development of proglacial lakes. When advancing downslope, away from a drainage divide, meltwater can drain freely from the ice margin and glaciofluvial outwash sediments may be deposited. Drainage divides that were significant regionally in determining the distribution of Quaternary sediments in Ontario were the Mississippi River–Great Lakes–St. Lawrence River drainage divide, the Mississippi River–Hudson Bay drainage divide and the Great Lakes–Hudson Bay drainage divide (*see* Figure 21.23). Other drainage divides were locally significant. These determined the location of outlet channels from the proglacial lakes (Figure 21.24).

The thickness of Quaternary sediments is quite variable across Ontario. In general, Quaternary sediments are very thin (less than 1 m) over large expanses of the rocky uplands of the Canadian Shield and along the bedrock escarpments in the lowland areas. Sediment or drift thickness is generally much greater in the Hudson Bay and Great Lakes–St. Lawrence lowlands where the thickness of drift can exceed 200 m, but is commonly between 30 and 60 m (Karrow 1989). Drift can also be thick, locally, along bedrock or structurally controlled valleys within the uplands of the Canadian Shield and may exceed 100 m in thickness (Dredge and Cowan 1989).

## General Characteristics of Ontario Tills

Till is the most widespread deposit left by the Laurentide Ice Sheet. It can be found as ground moraine in almost every area of the province. Till is, however, usually thin and discontinuous in upland areas of the Canadian Shield or along the crests of the large, bedrock-controlled escarpments in the Great Lakes–St. Lawrence Lowlands.

The general appearance of till depends on the mode of glacial transport, the depositional environment and processes and the source of the glacial debris. Post-depositional weathering processes can also affect till appearance and some of its characteristics.

Some of the effects that glacial transport and depositional process and environment have on the appearance of till were discussed in the section entitled Glacial Deposits and Landforms. The bedrock geology and preglacial landscape of Ontario produced 3 main end members of till: sandy tills derived from the erosion of Precambrian rocks, silty tills derived from the erosion of carbonate rocks, and clayey tills derived from the incorporation or deformation of

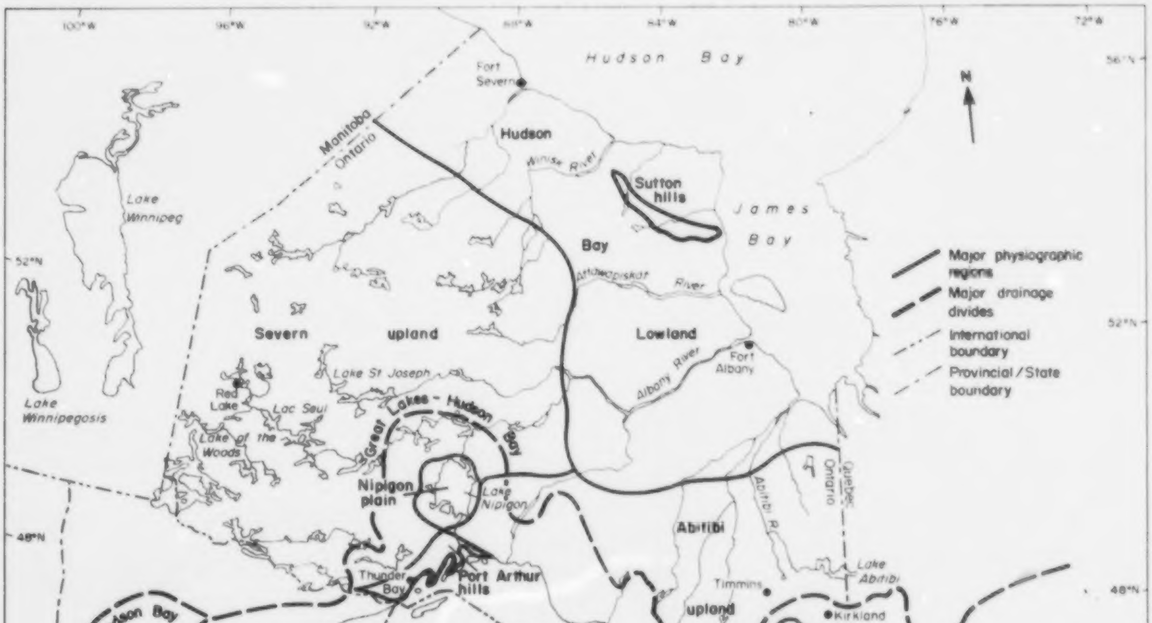

100°W
96°W
92°W
88°W
84°W
80°W
76°W
72°W
56°N
52°N
48°N
44°N
40°N
Hudson Bay
Fort Severn
N
Manitoba
Ontario
Hudson
Winisk River
Sutton hills
James Bay
Bay
Lowland
Attawapiskat River
Lake Winnipeg
Severn
upland
Lake St Joseph
Albany River
Fort Albany
Red Lake
Loc Seul
Lake Winnipegosis
Lake of the Woods
Great Lakes - Hudson Bay
Nipigon plain
Lake Nipigon
Ontario
Quebec
Abitibi R.
Abitibi
Thunder Bay
Port Arthur hills
Timmins
Lake Abitibi
upland
Kirkland Lake
Cobalt
Cobalt plain
Lake
Superior
Mississippi River - Hudson Bay
Sault Ste. Marie
Penokean hills
Sudbury
Ottawa River
St. Lawrence Lowlands
Lake Nipissing
Laurentian
Highlands
Ottawa
St. Lawrence R.
Georgian Bay
U S A
Mississippi
Lake Huron
Lake Michigan
U S A
Lake Simcoe
Toronto
Lake Ontario
St. Lawrence
Lowlands
London
U S A
Lake St. Clair
Windsor
Lake Erie
Missouri River
River
Mississippi River
Great Lakes
Ohio River
0 100 200 km
92°W
88°W
84°W
80°W
76°W
Major physiographic regions
Major drainage divides
International boundary
Provincial / State boundary

**Figure 21.23.** Major physiographic regions and major drainage divides delineated only in Ontario (*modified from* Bostock 1970a; Teller 1989).

fine-grained glaciolacustrine sediments that were deposited in front of the glacier in local or regional areas of ponding.

Till derived from erosion of the Precambrian igneous and metamorphic rocks of the Canadian Shield is commonly sandy textured (map unit 18). On average, sand constitutes over 70% of the till matrix, with usually less than 5% clay-sized particles. Clast content is usually high, with an abundance of crystalline rock types. The till is essentially noncalcareous, with matrix carbonate contents usually less than 5%.

The till derived from Precambrian rocks is usually nonplastic and commonly poorly graded as a result of the low percentage of silt- and clay-sized particles. This poor grading commonly results in the till being "loose", i.e., disaggregated within the weathering zone. The action of freeze-thaw cycles destroys any of the compaction or overconsolidation that may have developed during subglacial deposition. In addition, the percolation of surface waters down into the till is enhanced by the freeze-thaw cycles, which is in turn enhanced by the greater penetration of the surface waters. At depth, however, these sandy tills are often compact and display fissile structure. They have commonly been thought of as being 2 different tills because of their differences in compaction; however, this is not usually the case.

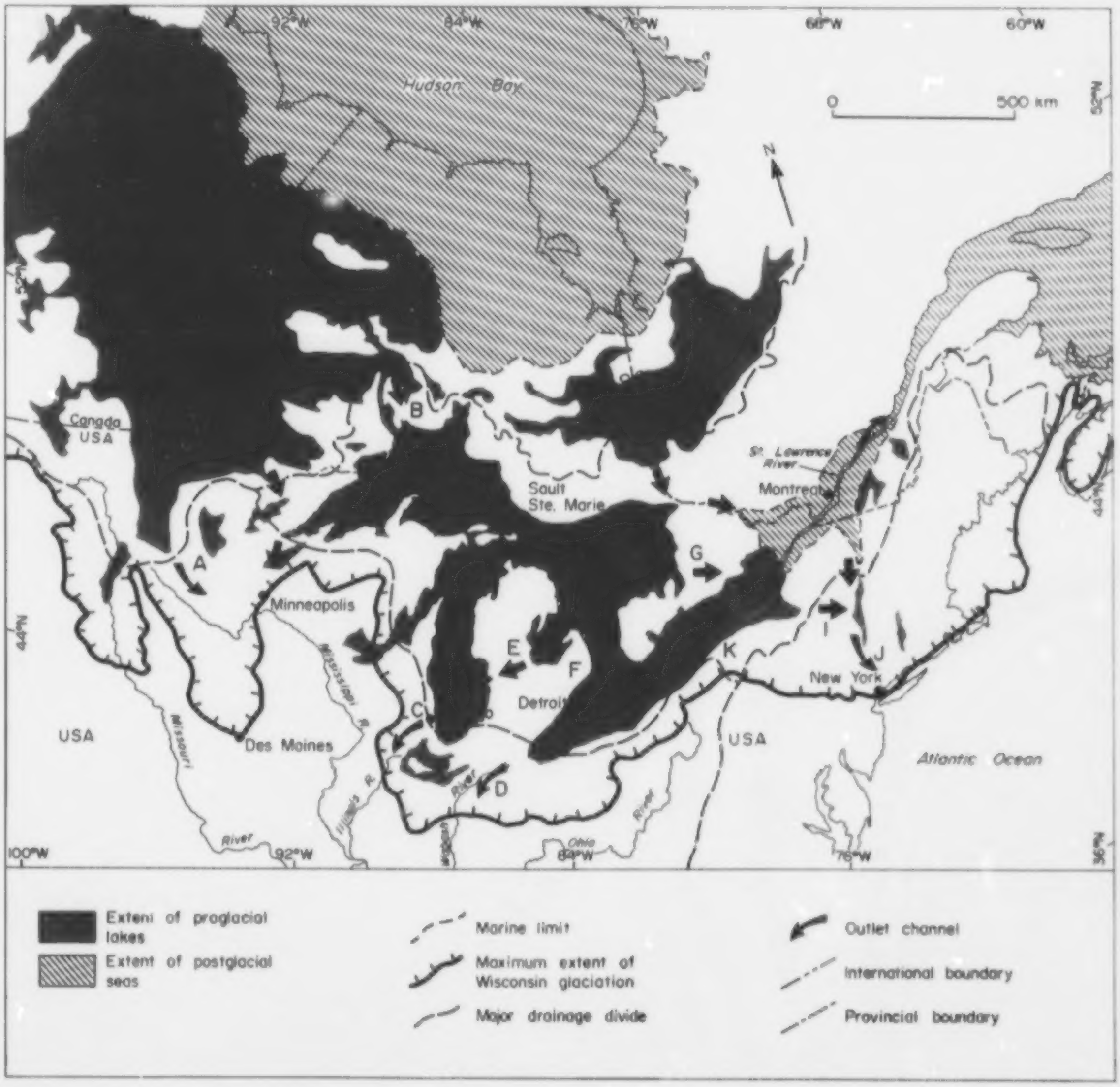

92°W
84°W
76°W
68°W
60°W
Hudson Bay
0
500 km
52°N
N
Canada
USA
B
Sault Ste. Marie
St. Lawrence River
Montreal
G
A
Minneapolis
Mississippi R.
E
F
K
I
J
New York
Detroit
C
USA
USA
Missouri
Des Moines
Illinois R.
Wabash River
D
Ohio River
River
River
Atlantic Ocean
44°N
100°W
92°W
84°W
76°W
36°N
Extent of proglacial lakes
Extent of postglacial seas
Marine limit
Maximum extent of Wisconsin glaciation
Major drainage divide
Outlet channel
International boundary
Provincial boundary

**Figure 21.24.** Outlet channels and areas of ponding. Major outlet channels are marked with arrows and labelled: A - Minnesota River Valley; B - Eastern Agassiz Outlets; C - Chicago or Des Plaines Outlet; D - Wabash River Valley; E - Grand River Valley; F - Port Huron Outlet; G - Kirkfield Outlet or Fenelon Falls; H - North Bay Outlet; I - Rome, New York or Mohawk River Valley; J - Hudson River Valley; K - Buffalo or Niagara River Outlet (*modified from* Teller 1989).

In areas where the till cover over the rock surface is very thin, as occurs within map unit 1, subglacially deposited till is typically very stony, loose and sandy, and has frequently been mistaken for supraglacial (ablation) till. Stoniness in tills derived from Precambrian rock appears to have an inverse relation to till thickness (Figures 21.25 and 21.26).

Sand layers and lenses are common in lodgement till deposited on the Precambrian Shield. The irregular bedrock surface and large boulders being transported along the base of the glacier create cavities under the ice into which meltwater can flow. The fine component of the coarse-textured glacial debris, which has fallen, melted out or flowed into the subglacial cavities, is easily winnowed away, producing discontinuous layers or lenses of gravelly sand and sand. These sand layers can be distinguished from those found in melt-out till because they are usually deformed by a unidirectional applied stress and are commonly attenuated in the down-ice direction. Sand stringers are abundant in subglacial till deposited against steep, northward-facing bedrock surface slopes, probably as a result of opening and closing of cavities combined with subsole and internal folding and faulting in response to local compressive flow conditions. Multiple till layers, separated by usually thin beds of stratified sediments, can be generated on the gentle lee sides of rock ridges as the result of sedimentation in cavities that were created under extending glacier flow conditions (Sugden and John 1976).

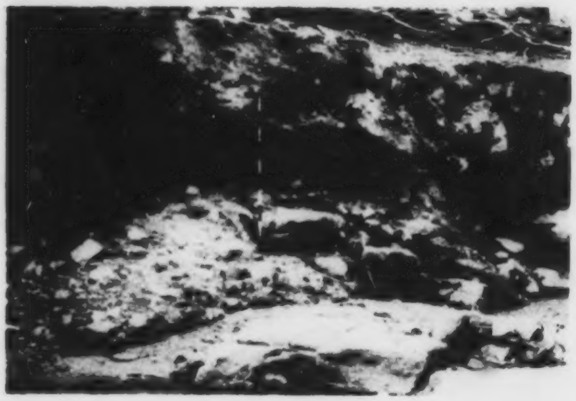


**Figure 21.25.** Subglacial till, lodgement facies near Baptiste Lake, about 15 km west of Bancroft. Lithologies of the clasts are predominantly locally derived (local till). Bedrock is exposed at base of photograph.

Within large cavities formed in the lee of bedrock knobs, thick interbedded sequences of sand and diamicton beds of both flow and melt-out origin can be found. These commonly contain glaciotectonic structures formed during the closing of the cavity following the major melt season (Figure 21.27).

The types of Precambrian rocks immediately up-ice from the till observation site greatly affect the overall appearance of the till, and in certain instances its texture. In

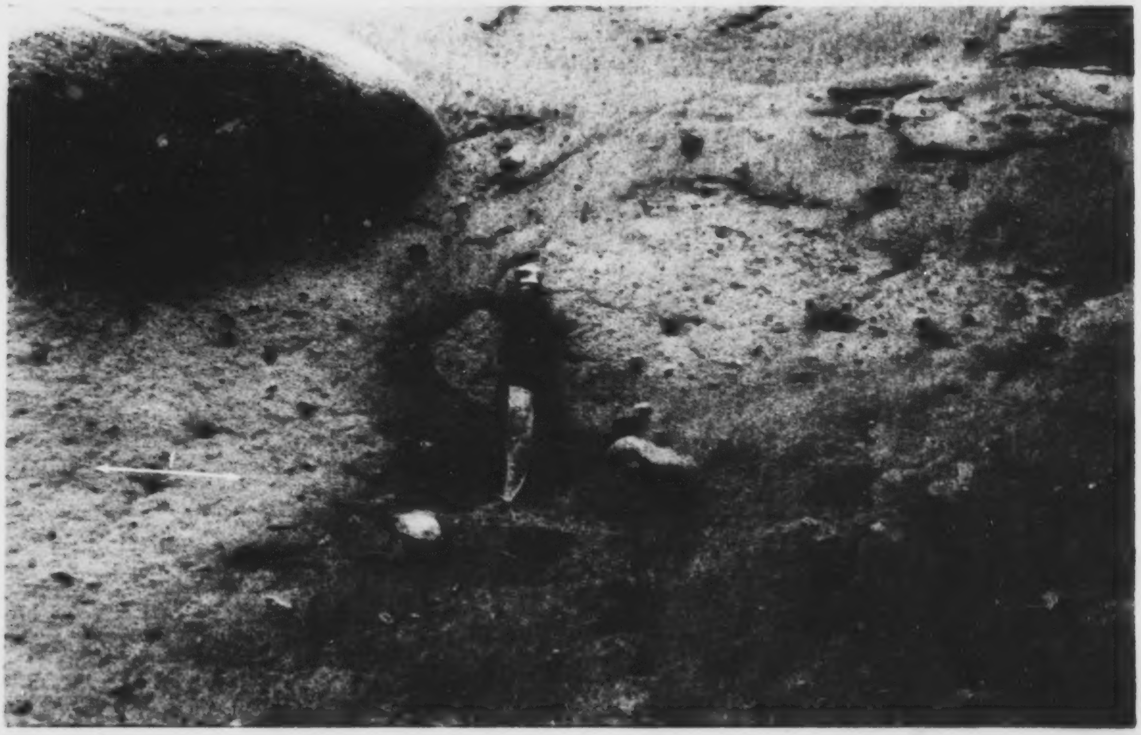


**Figure 21.26.** Subglacial till, lodgement facies in drumlin, Maynooth, approximately 25 km north-northwest of Bancroft, showing well-developed clast fabric and fissility (f). (Knife is about 20 cm long.)

general, easily eroded mafic rocks produce finer-textured tills than tills derived from felsic rocks. The type of Precambrian rock from which the till is derived can affect the productivity of soils developed on the till.

Till derived from the erosion of carbonate rocks, limestones and dolostones is usually silty textured (map unit 19 and thin in areas of map unit 2). Matrix grain-size distribution is usually between 25 to 40% sand, 40 to 50% silt and 15 to 30% clay. Up to 80% carbonate clasts have been reported (Dredge and Cowan 1989). Matrix carbonate content is high, commonly between 30 and 50%.

Till derived from carbonate rocks is nonplastic and usually well graded. The well-graded nature allows for better packing of grains, and usually results in compact to very compact tills with preservation of the original overconsolidation, produced by glacial loading during deposition. A carbonate cement, produced during the weathering of the carbonate-rich matrix and clasts, may also add to the strength of this material. For example, a till exposed in the Uhtoff Quarry near Orillia has already become lithified (Figure 21.28).

Carbonate-derived tills are more massive in appearance than those deposited on the Precambrian Shield, except in the marginal zones where the effects of depositional processes become dominant. The typically smoothed, stepped bedrock surfaces associated with the Paleozoic carbonate rocks in Ontario are probably the reason or at least a contributing factor. Fissility is a common structure found in tills derived from the erosion of carbonate rocks. Produced by the release of stress, fissility may reflect the original increments of deposition and shear during the lodgement process.

Fine-grained tills form the final end member of the 3 basic types of tills in Ontario (map unit 21). Such tills formed as the result of the incorporation of fine-grained glaciolacustrine sediments into the base of the ice prior to deposition or by the subsole deformation of this fine-grained material. These tills can contain up to 55% clay-sized particles (usually between 30 and 40%), 40 to 70% silt, with sand content usually below 15%. The clast content in these fine-grained tills is usually low. It has been suggested that for every metre of fine-grained Port Stanley Till deposited in the Lake Erie basin, up to 0.9 m of glaciolacustrine sediment went into its production (Barnett 1987).

Subglacial varieties of these fine-grained tills are commonly massive and homogeneous in appearance and may have a large blocky structure or contain vertical jointing (Figure 21.29). Although subglacially deposited, these fine-grained tills are not usually highly overconsolidated. Being deposited by thin, fast-flowing ice lobes in an environment with ponded water at the front of the glacier (usually in lake basins), the rate of the expulsion of porewater within the fine-grained, subglacially deposited debris was probably slow. A substantial amount of the glacial load

**Figure 21.27.** Subglacial till deposited in lee of bedrock obstacle, near New Heron, approximately 20 km east of Bancroft. Glacial debris and sands resulting from the washing of glacial debris in a cavity, which formed at the base of the glacier in the lee of a bedrock knob, are sheared and attenuated in the down-ice direction by subsole drag.

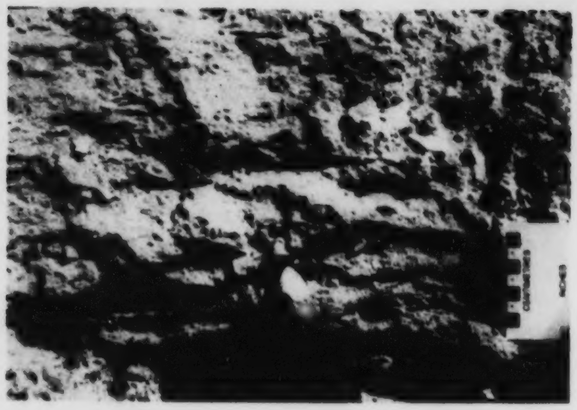


**Figure 21.28.** Lithified subglacial lodgement till, Uhtoff, about 15 km northwest of Orillia.

**Figure 21.29.** Clay-rich till exposed along the Lake Erie shore bluffs, 5 km west of Port Burwell.

may have been removed before porewater pressures were dissipated and consolidation could begin (Barnett 1987).

A general coarsening of fine-grained till occurs towards the margins of lake basins through the incorporation of coarser-textured glaciolacustrine sediments deposited along the lake edge. The Wentworth Till provides a good example (Figure 21.30).

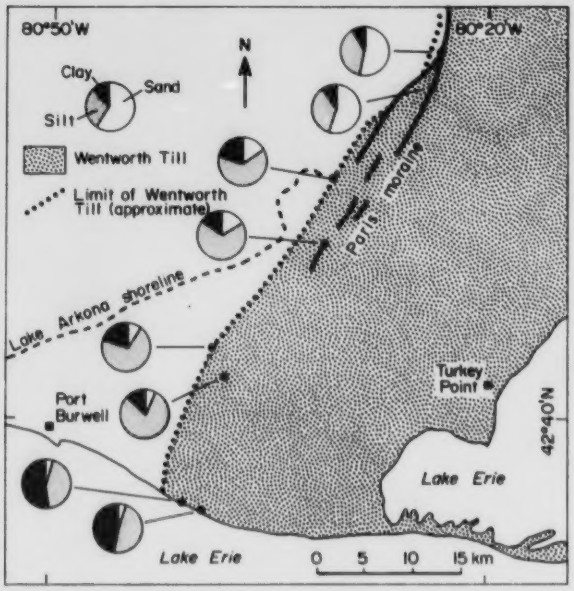


**Figure 21.30.** Grain-size variation in the Wentworth Till as the result of the incorporation of glaciolacustrine sediments (*modified from* Sharpe and Barnett 1985).

The 3 general types of tills described above are end members of the continuous spectrum of tills found in Ontario. The ice centres of the Laurentian Ice Sheet in Keewatin and Labrador overlie Precambrian rocks of the Canadian Shield. The Laurentide Ice Sheet, enroute to its southern limit in the northern United States during the Late Wisconsinan, traversed sedimentary rocks of the Hudson Bay Lowland, Precambrian rocks of the southern Canadian Shield and finally sedimentary rocks of the Great Lakes–St. Lawrence Lowlands. Dispersal of carbonate clasts and matrix carbonate in tills on the northern edge of the Canadian Shield, down-ice from the carbonate rocks of the Hudson Bay Lowland, has been substantial (Figure 21.31). In a like manner, debris originating from the Precambrian Shield was dispersed across the Great Lakes–St. Lawrence Lowlands. The various debris types were in places mixed with proglacial lake sediments and pre-existing sediments to produce the mélange of tills that are found across Ontario.

Within the Great Lakes–St. Lawrence Lowlands, numerous till sheets were deposited during the advance and retreat of the Laurentide Ice Sheet margin. Ice flow was controlled by the broad topographic depressions of the Great Lakes basins. Lobes of ice extending beyond the main body of the ice sheet developed in these basins, and at times, acted independently in response to local conditions at the base of the glacier rather than, or in addition to, climatic change. The individual lobes are referred to by the name of the lake basin or basins in which they are located (Figure 21.32).

The tills within the Great Lakes area occur commonly in sheets 2 to 10 m thick; most of which have been traced laterally for tens of kilometres. These till sheets can contain all the various genetic varieties of till, particularly along former ice-marginal positions where end or interlobate moraines have formed.

# Stratigraphy of Ontario's Sediment Record

## INTRODUCTION

The stratigraphic framework (*see* Figure 21.4) and nomenclature of Quaternary sediments in Ontario have been summarized by Dreimanis and Karrow (1972), Karrow (1974, 1984, 1987, 1989) and Cowan (Cowan et al. 1975, 1978; Dredge and Cowan 1989). There is a much better understanding of the distribution of the Quaternary sediments deposited during the Late Wisconsinan than of those deposited earlier. Exposures of the older deposits are rare and commonly deeply buried beneath the Late Wisconsinan sediments. Interpretations of the origin and depositional environments of these older sediments are difficult and are usually based on limited evidence and exposure. Correlations are even more difficult, for they must be made over very large distances. In addition, reliable absolute dating methods for sediments deposited prior to the Late Wisconsinan are wanting. Nevertheless, useful attempts to classify and correlate these older sediments have been made, as described below.

## PRE–LATE WISCONSINAN RECORD

Exposures of sediments deposited prior to the last major glacial advance or before the Late Wisconsinan substage (oxygen isotope stage 2) are rare. They are scattered throughout the province, generally occurring in areas of relatively thick glacial drift. These pre–Late Wisconsinan deposits occur for the most part at depth and are only observed in natural exposures such as riverbanks or lake bluffs, or in man-made exposures such as sand and gravel pits, quarries and mines. Other occurrences have been reported from temporary road excavations and from borehole cores and cuttings. Natural sites occur in the Hudson Bay Lowland (Dredge and Cowan 1989) and in the interlake peninsula of southern Ontario (Karrow 1989). Man-made sites or boreholes encountering pre–Late Wisconsinan sediments occur in the clay belts of Timmins (Brereton and Elson 1979) and Fort Frances (A.F. Bajc, Ontario Geological Survey, personal communication, 1990), along the Ottawa River valley and the interlake peninsula of southern Ontario (Karrow 1989).

### Illinoian Glacial Record

The oldest deposits found in Ontario were deposited during or possibly before the Illinoian Glaciation, prior to 135 000 years ago (*see* Figure 21.4). These deposits include the York Till and older tills in the Toronto area, 3 or more till layers in the Hudson Bay Lowland, and possibly the Bradtville till encountered in boreholes along the north shore of Lake Erie. An old till reported in boreholes in the Fort Frances area may also be Illinoian (A.F. Bajc, Ontario Geological Survey, personal communication, 1990). The older tills in the Toronto area and in the Hudson Bay area underlie nonglacial deposits containing fossil evidence of a climate as warm or warmer than today; therefore, they predate the last interglacial deposits of the Sangamonian (oxygen isotope stage 5e). Sangamonian deposits appear to be absent in the Fort Frances area and along the north shore of Lake Erie, so these tills may be younger and deposited during glacial advances in the Early or Middle Wisconsinan.

The York Till (Terasmae 1960) has been observed in excavations in downtown Toronto and at the Don Valley Brickyards overlying Ordovician shale bedrock, but is best exposed near Woodbridge, Ontario (Figure 21.33). It is a clayey sand till which contains numerous shale and limestone clasts derived from local bedrock sources (Karrow 1967). Striations on the bedrock surface beneath this till indicate that ice flow was to the north out of the Lake Ontario basin (Terasmae 1960). Boreholes in the vicinity of the Woodbridge cut have encountered another layer of till beneath the York Till.

In the Hudson Bay Lowland, Terasmae and Hughes (1960a, 1960b) and Skinner (1973) reported the presence of 3 layers of till beneath the interglacial Missinaibi Formation at Site 24M along the Missinaibi River. Shilts (1984a) subsequently identified a fourth till layer directly below the oldest till of Skinner's. The characteristics of these 4 till layers are quite similar. Texturally, they are all stony sand tills. Sands and gravels, containing sedimentary structures which indicate southward current flows during deposition, and laminated silts and clays occur between the upper till layers. The sequence and characteristics of the sediments led Skinner to conclude that the layers of till were deposited by an oscillating ice front in a proglacial lake and represent only one main glaciation. Shilts (1984a), however, noting the extent of oxidation in the gravels, suggested that a major nonglacial interval is recorded in the sediment sequence beneath the Missinaibi Formation (Dredge and Cowan 1989). Most of the till layers below the Missinaibi beds appear to be made up of flowtills and the oxidation of the gravels is probably the result of groundwater movements. The sediment sequence beneath the Missinaibi Formation appears to reflect a former ice-marginal position in a proglacial lake, dominated by sediment gravity flows or flowtills, and like Skinner's interpretation, only one glaciation is needed.

Other exposures of till and intertill sediments, believed to be older than the Missinaibi Formation, are found in riverbank exposures elsewhere in the Moose River drainage basin. Tills beneath organic-bearing sediments in the Timmins–Kapuskasing area may also be Illinoian or older (R.N.W. DiLabio and S.L. Smith, Geological Survey of Canada, personal communications, 1989). Further work is needed in these areas to unravel this complex Illinoian record.

### Sangamonian Interglacial Record

The Sangamonian Interglacial is represented in the Hudson Bay Lowland area by the Missinaibi Formation (McDonald

1969; Skinner 1973), in the Toronto area by the Don Formation (Coleman 1894) and possibly in the Timmins area by the Owl Creek beds (R.N.W. DiLabio, Geological Survey of Canada, personal communication, 1990). The fossils in all of these deposits indicate climatic conditions similar to that of today or warmer than today, hence, the interglacial interpretation. Organic-bearing sediments at depth from beneath the pit floor at the Point Fortune Site in eastern Ontario may also have been deposited during the Sangamonian. Fossil assemblages suggest that a climate warmer than today existed during deposition of the organic-bearing strata (T.W. Anderson, J.V. Matthews and R.J. Mott, Geological Survey of Canada, personal communications, 1987).

The Don Formation overlies the Illinoian York Till. It consists of fossiliferous gravels, sands, silts and clays interpreted as being deposited at the mouth of a large river in a lake embayment (Williams and Morgan 1977). The lake level in which these sediments were deposited was several metres higher than the present level of Lake Ontario.

Numerous types of plant and animal remains in Don Formation sediments indicate a warm climate up to 3°C warmer than today (Kelly et al. 1987; Karrow 1989). An upward cooling trend through the Don Formation is indicated by the fossils. The lowermost beds may, however, represent the warmest part of the Sangamonian that was deposited about 125 000 years ago.

The Missinaibi Formation as defined by Skinner (1973; McDonald 1969) contains a lower marine member, a fluvial member, a forest bed member (Figure 21.34) and an upper glaciolacustrine member. This sequence of sediments is similar to sediments deposited following the last glaciation (McDonald 1969). The isostatically depressed land surface became inundated by the sea initially after deglaciation. This was followed by uplift, fluvial downcutting and sedimentation (with direction of streamflow toward Hudson Bay); the result of isostatic rebound. Once exposed, the landscape became vegetated as it has today. With a subsequent glacial advance, however, proglacial lakes would form between the advancing ice sheet and the Great Lakes–Hudson Bay drainage divide to the south, explaining the deposition of glaciolacustrine sediments over the forest bed.

Fossil remains indicate climate was as warm as today or possibly warmer and the sediment sequences indicate that substantial isostatic recovery had occurred (Dredge and Cowan 1989). Both of these lines of evidence support an interglacial rank for the Missinaibi Formation.

The Owl Creek beds are predominantly organic-bearing silts with pollen assemblages indicating conditions similar to today. They occur stratigraphically beneath the Matheson till and were deposited during the Sangamonian Interglacial (R.N.W. DiLabio and R.J. Mott, Geological Survey of Canada, personal communications, 1989).

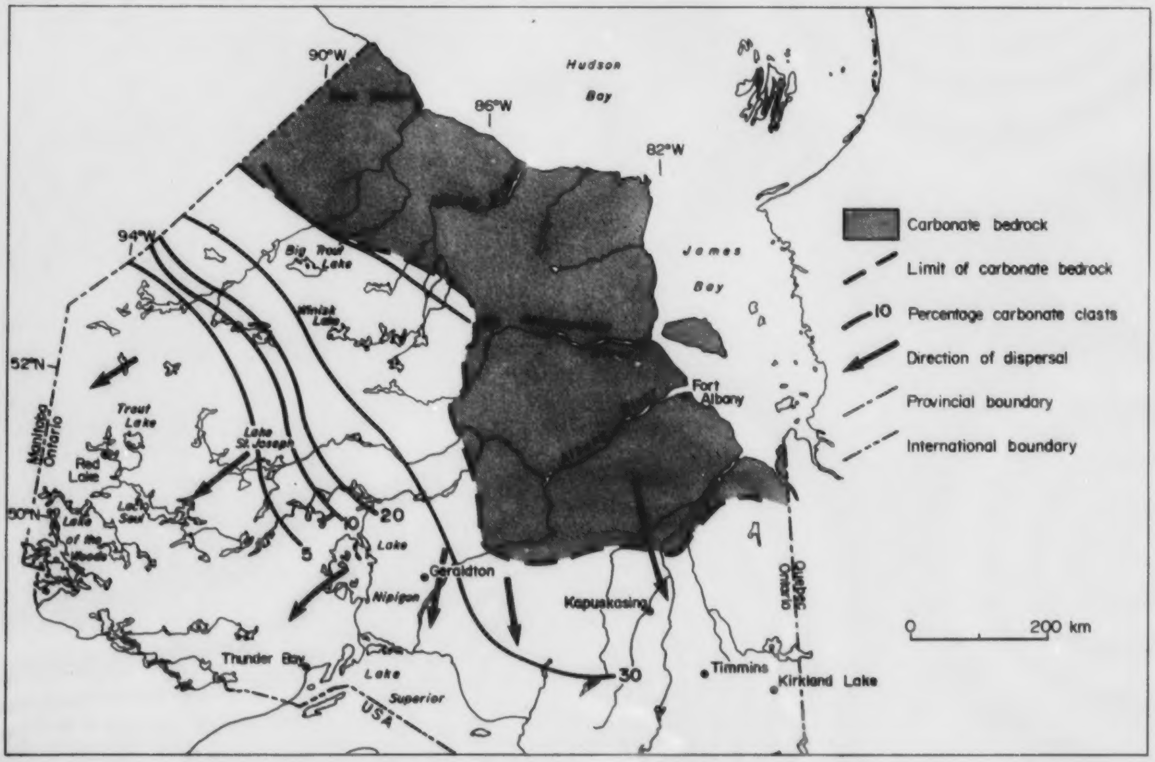

90°W
86°W
82°W
94°W
52°N
50°N
Hudson Bay
James Bay
Big Trout Lake
Winisk Lake
Trout Lake
Lake St. Joseph
Red Lake
Lac Seul
Lake of the Woods
Manitoba
Ontario
Lake Nipigon
Geraldton
Thunder Bay
Lake Superior
USA
Kapuskasing
Timmins
Kirkland Lake
Fort Albany
Quebec
Ontario
A
5
10
20
30
Carbonate bedrock
Limit of carbonate bedrock
10 Percentage carbonate clasts
Direction of dispersal
Provincial boundary
International boundary
0
200 km

**Figure 21.31.** Dispersal of carbonate rock fragments in tills of northwestern Ontario: a) dispersal of carbonate clasts; b) dispersal of carbonate minerals within the till matrix. Arrows indicate direction of dispersal (*modified from* Dredge and Cowan 1989).

## Early Wisconsinan Record

The most complete and best documented record of the Early Wisconsinan is from the Toronto area (*see* Figure 21.33). Here, 3 stratigraphic units have been placed within the Early Wisconsinan; the Scarborough Formation, Pottery Road Formation and the Sunnybrook drift.

The Scarborough Formation (Karrow 1967), well exposed along the lower half of the Scarborough Bluffs, consists of approximately 50 m of a gradually coarsening upward sequence beginning as clay and silt rhythmites at the base, and becoming channelized cross-bedded sands at the top (Kelly and Martini 1986). It is interpreted as a delta at the mouth of a large river which flowed into glacial Lake Scarborough (Coleman 1941). Glacial Lake Scarborough had a water level similar to the postglacial Lake Iroquois, about 45 m above the present level of Lake Ontario. In order to have a lake at this level in the Lake Ontario basin, an ice sheet must have blocked the St. Lawrence River valley, diverting flow to the Hudson River drainage system via the Mohawk River valley in New York State (*see* Figure 21.24; Karrow 1967, 1989).

Studies of pollen and macrofossils indicate that the climatic conditions during deposition of the Scarborough Formation were cool to cold, and a northern boreal forest to treeline environment probably existed in the Toronto area (Terasmae 1960; Morgan 1972, 1975; Williams et al. 1981). The Scarborough Formation is believed to have been deposited during the Nicolet Stade between 106 000 to 115 000 years ago (*see* Figures 21.4 and 21.35).

The Pottery Road Formation, consisting of weakly fossiliferous gravels and gravelly sands, occurs only within large and small channels cut into the Scarborough and Don formations (Karrow 1974). Vertebrate and mollusc remains in these sediments provide little evidence of climatic conditions. These high-energy channelized gravel and gravelly sand deposits have been interpreted as river alluvium graded to a lower lake level in the Lake Ontario basin (Karrow 1974, 1989). This would require the withdrawal of the ice sheet margin from the St. Lawrence River valley. The Pottery Road Formation is commonly correlated with the cool climate, nonglacial St. Pierre beds deposited during the St. Pierre Interstade, some 75 000 to 106 000 years ago (*see* Figure 21.4; Karrow 1989).

Alternatively, channel erosion and sediment infilling that have been assigned to the Pottery Road Formation in the Don Valley Brickyards may have been cut by meltwater and filled proglacially or subglacially (Sharpe and Barnett 1985). If the other channels of this formation were cut by meltwater, the correlation with the St. Pierre beds would be unlikely.

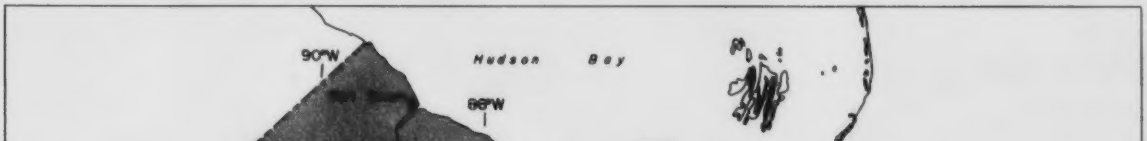

Hudson Bay
90°W
86°W
82°W
94°W
52°N
50°N
James Bay
Fort Albany
Trout Lake
Winisk Lake
Trout Lake
Lake St. Joseph
Red Lake
Lac Seul
Lake of the Woods
Lake Nipigon
Geraldton
Kapuskasing
Timmins
Kirkland Lake
Thunder Bay
Lake Superior
USA
Manitoba
Ontario
Quebec
Ontario
30
10
0
200 km
Carbonate bedrock
Limit of carbonate bedrock
10 Percentage of carbonate mineral
Direction of dispersal
Provincial boundary
International boundary

**Figure 21.31.** Dispersal of carbonate rock fragments in tills of northwestern Ontario: a) dispersal of carbonate clasts; b) dispersal of carbonate minerals within the till matrix. Arrows indicate direction of dispersal (*modified from* Dredge and Cowan 1989).

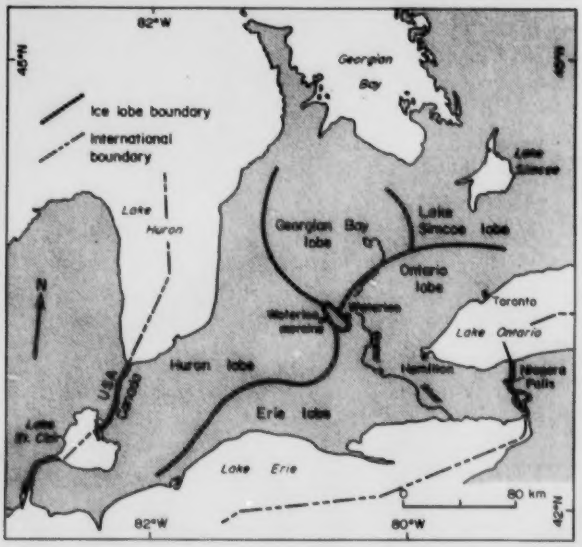


**Figure 21.32.** Influence zones of the major ice lobes of the lower Great Lakes basins (*modified from* Cowan et al. 1978).

The Sunnybrook drift (Karrow 1969) overlies the Pottery Road and Scarborough formations in the Toronto area. The Sunnybrook drift is composed of a lower member termed the Sunnybrook Till (Terasmae 1960), and an associated upper member, varved clay unit termed the Bloor Member. The Sunnybrook Till varies in thickness between 6 and 10 m and is a fine-grained silt to silty-clay till. Clast content is low, but contains a notable content of Potsdam sandstone, which outcrops only to the east of Toronto. The Sunnybrook Till is generally believed to have been deposited by an ice lobe flowing radially out of the Lake Ontario basin (Dreimanis and Terasmae 1958; Karrow 1967).

The Sunnybrook Till is not a single massive till layer, but is made up of multiple massive diamicton (till-like) beds. The multiple unit aspect, combined with contact relationships, led Eyles and Eyles (1983) to conclude that the Sunnybrook Till was not a till at all, but a lake deposit. The discontinuous stratification within the Sunnybrook Till, however, is probably the result of interfingering layers of ice-marginal flowtills and subglacial deformation and lodgement till (Sharpe and Barnett 1985; Barnett 1987; Hicock and Dreimanis 1989).

The Sunnybrook drift represents an advance of the continental ice sheet at least as far south as Toronto during the Guildwood Stade, 60 000 to 75 000 years ago.

## Middle Wisconsinan Record

Middle Wisconsinan deposits occur widely throughout southwestern Ontario. They usually occur as thin, discontinuous sheets or lenses of nonglacial sediments that were deposited in small lakes, ponds or along river and stream valleys (Karrow 1989).

Middle Wisconsinan deposits in the Lake Erie basin consist of a buried soil (Port Talbot I Interstade) and 2 main organic-bearing units (Port Talbot II and Plum Point interstades), each separated by glaciolacustrine fine-grained sediments (Dreimanis et al. 1966; Barnett et al. 1987). The presence of a continental ice sheet margin in the eastern end of the Lake Erie basin, blocking the Buffalo outlet (*see* Figure 21.24), is required at least once, possibly twice, during the Middle Wisconsinan for the formation of the lakes into which these intervening glaciolacustrine sediments were deposited.

Karrow (1989) has summarized most of the Middle Wisconsinan sites in southern Ontario. Cool climatic conditions are indicated by the fossils contained in these scattered sites. Boreal forest-type vegetation probably covered the southern Ontario landscape during this time interval (Karrow 1989).

In the Toronto area, the Middle Wisconsinan is represented by the Thorncliffe Formation (*see* Figure 21.33; Karrow 1967). This formation is composed of stratified sediments of sand, silt and clay of both glaciofluvial and glaciolacustrine origin. Two layers of till along the Lake Ontario shore, the silty Seminary Till and the clayey Meadowcliffe Till, divide this formation into 3 members (Karrow 1989). Farther inland, the 2 till layers appear to pinch out and the Thorncliffe Formation is 1 continuous unit. The glacial origin of the Seminary and Meadowcliffe tills has also been questioned (Eyles and Eyles 1983).

The elevation, origin and distribution of the Thorncliffe Formation sediments indicate that drainage down the St. Lawrence River valley was once again blocked. The ice sheet margin was within the Lake Ontario basin, and if the interpretation that the Seminary and Meadowcliffe tills are tills is correct, then the ice margin reached south to at least Toronto. The Middle Wisconsinan sediment sequences in the Lake Erie basin also require an ice sheet margin at least in the western end of the Lake Ontario basin.

In the Hudson Bay Lowland, the record of the Middle Wisconsinan is confusing at best. As many as 4 till layers are believed to exist, usually separated by thin beds of stratified sediments (Dredge and Cowan 1989). In the Moose River drainage basin, McDonald (1969) and Skinner (1973) recognized only 2 layers of till (the Adam and Kipling tills) separated by cross-stratified sands, and silt and clay rhythmites which Skinner referred to as the "Friday Creek sediments". Skinner (1973) believed that if the Missinaibi Formation was an interglacial deposit, then there was probably continuous ice cover over the Moose River drainage basin during the entire Wisconsinan. Dredge and Nielsen (1985) and Dredge and Cowan (1989) agreed with a continual ice cover over the Hudson Bay Lowland during the Wisconsinan, and suggested a subglacial origin for the intertill stratified sediments.

Andrews et al. (1983) suggested that the Hudson Bay Lowland was not continuously covered by a continental ice sheet during the Wisconsinan. Based on the grouping of amino acid ratios of shells collected from till and intertill sediments, Andrews (1989; Andrews et al. 1983) suggested that the lowland was ice-free twice, approximately 74 000

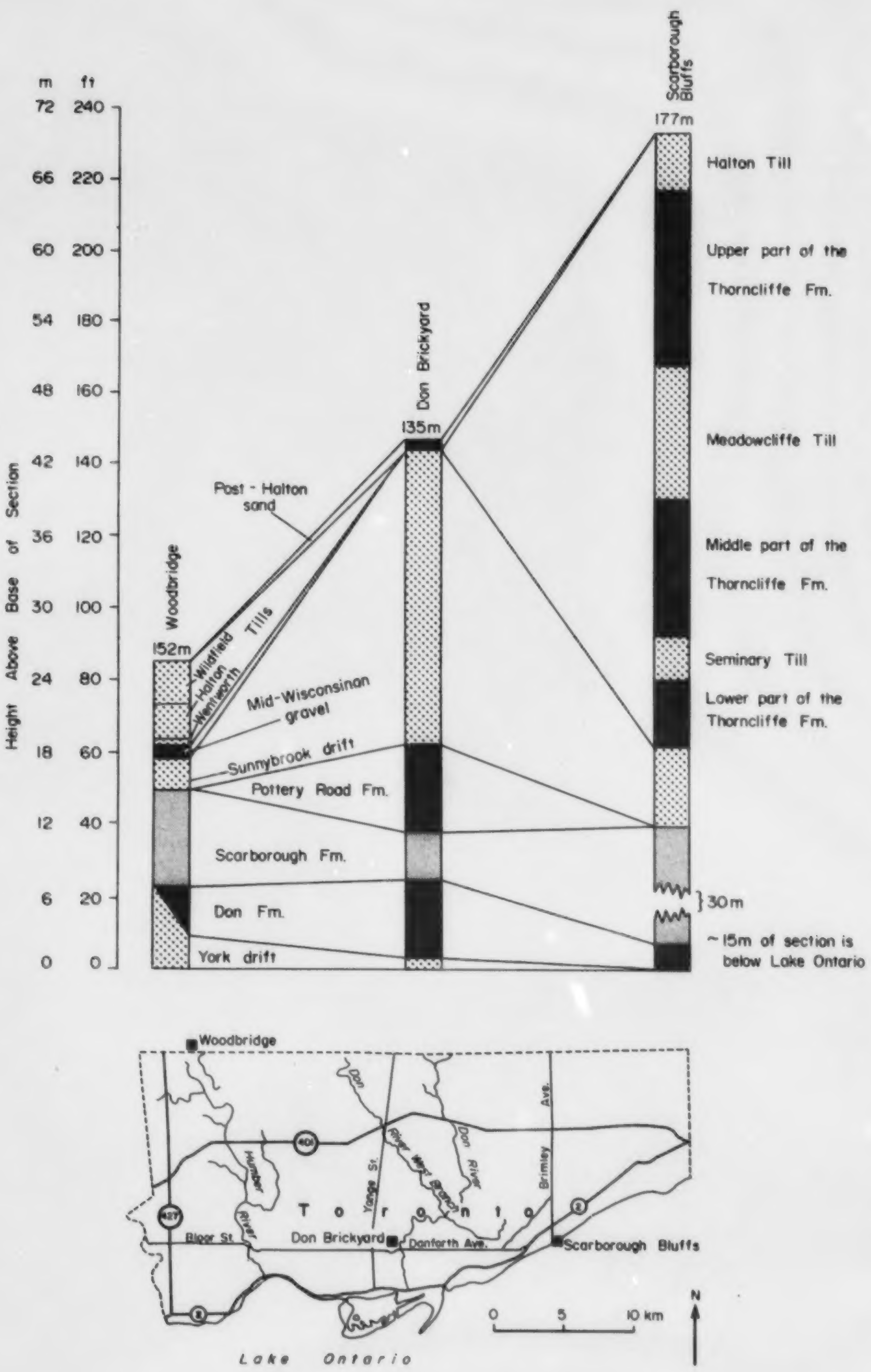

m
ft
72 240
66 220
60 200
54 180
48 160
42 140
36 120
30 100
24 80
18 60
12 40
6 20
0 0
Height Above Base of Section
Woodbridge
152m
Don Brickyard
135m
Scarborough Bluffs
177m
Post - Halton sand
Wildfield
Halton
Wentworth
Tills
Mid-Wisconsinan gravel
Sunnybrook drift
Pottery Road Fm.
Scarborough Fm.
Don Fm.
York drift
Halton Till
Upper part of the Thorncliffe Fm.
Meadowcliffe Till
Middle part of the Thorncliffe Fm.
Seminary Till
Lower part of the Thorncliffe Fm.
30m
~15m of section is below Lake Ontario
Woodbridge
Don
401
Humber
Yonge St.
Don River West Branch
Don River
Brimley Ave.
427
T o r o n t o
2
River
Bloor St.
Don Brickyard
Danforth Ave.
Scarborough Bluffs
2
0 5 10 km
N
Lake Ontario

**Figure 21.33.** Important stratigraphic sections in Toronto containing pre–Late Wisconsinan sediments (*modified from* Karrow 1984). Numbers at the top of each section are elevations above sea level.

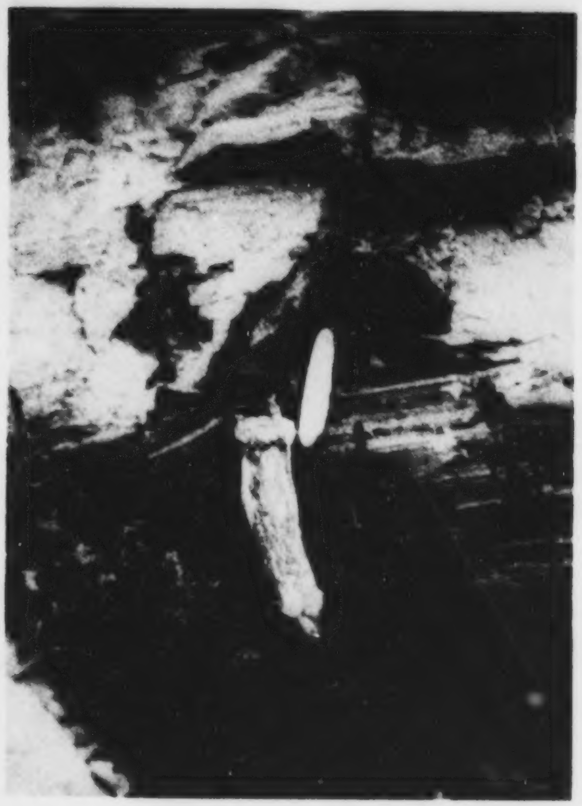


**Figure 21.34.** Forest bed member of the Missinaibi Formation as exposed along Adam Creek diversion channel, about 45 km north of Smooth Rock Falls. The tree stump is in growth position. Note the bulge at the top of the stump caused by downward pressure exerted during the overriding of the site by the Laurentide Ice Sheet (photograph *courtesy of* E.V. Sado). (Knife is about 8 cm long.)

and 36 000 years ago during both the Early and Middle Wisconsinan.

Amino acid dating is a relative dating method which measures the abundances and ratios of the various isomers of the amino acids present in a fossil material, such as shells or wood. The amino acid ratios are functions of temperature, time since death and genus. After death, however, the proteins (which are made up of amino acids) contained within the fossil material must be broken by hydrolysis before the racimization process, the changing from the original amino acid isomer to another, can begin.

Andrews (1989) suggested that because past temperature gradients in the Hudson Bay Lowland have been low and that the area is in a homogeneous climatic region, the differences in the amino acid ratios largely reflect age differences of the dated shells and the containing sediments.

The temperatures, however, that the shells were subjected to probably fluctuated above and below 0°C. Some of the shells spent various lengths of time being transported within the glacier at temperatures below which the hydrolysis process ceases. The amino acid ratios obtained, therefore, may be suspect, and the amino acid stratigraphy developed from them is questioned.

Much of northern Ontario was probably ice covered during the Early and Middle Wisconsinan. However, ice-marginal retreat into the lowlands should not be entirely ruled out. Further detailed sedimentological and stratigraphic studies are needed.

## LATE WISCONSINAN RECORD

In Ontario, the record of the Late Wisconsinan is by far the most extensive, continuous and best understood. Almost all parts of Ontario bear evidence of the last major ice advance that covered the entire province. The accompanying maps (*see* Maps 2553, 2554, 2555 and 2556, map case) display the distribution of the materials deposited by the Laurentide Ice Sheet during its final advance and retreat.

During Late Wisconsinan time, 3 significant periods of ice advance directly affected the lower Great Lakes region; they are termed the Nissouri, Port Bruce and Port Huron stades (Dreimanis and Karrow 1972). These stades were separated by 2 periods of general ice-margin recession or warmer periods, the Erie and Mackinaw interstades, and were followed by the Two Creeks Interstade (*see* Figures 21.4 and 21.35). The Greatlakean Stade is discussed in the section entitled Later Events and Sediments for its extent and significance in Ontario is questionable.

### Nissouri Stadial Sediments

The Nissouri Stade is the initial stage of ice advance during the Late Wisconsinan. During this stade, the Laurentide Ice Sheet reached the Niagara Escarpment by about 23 000 years ago (Hobson and Terasmae 1969), and its southernmost extent, marked by the Cuba and Hartwell moraines in Ohio and Indiana, by approximately 18 000 to 20 000 years ago (Goldthwait et al. 1965; Dreimanis and Goldthwait 1973).

The Nissouri Stadial in Ontario is represented by the Catfish Creek Drift (de Vries and Dreimanis 1960). Catfish Creek Drift outcrops along Catfish Creek near Sparta and in the Lake Erie bluffs in the vicinity of Port Talbot. Catfish Creek Drift is composed of several layers of subglacial till as well as stratified sediments of glaciofluvial and glaciolacustrine origin and supraglacial till layers and lenses.

The main layer of Catfish Creek Till (map unit 3) is widespread throughout southwestern Ontario in the subsurface, and outcrops in small areas in the vicinity of Woodstock and Dundalk (map unit 3). It is remarkable for its overall consistency in composition and has been used as a valuable marker bed for stratigraphic studies (Cowan 1978; Karrow 1988, 1989). Catfish Creek Till is a moderately stony to very stony, very compact, highly calcareous, sandy silt to silt till. Matrix carbonate content averages 35 to 60% and is mainly dolomitic. Clast content is usually between 10 and 25%. It is often referred to as "hardpan" in water well drillers' records because of its stiffness. Catfish Creek Till can be very difficult to excavate as a result of its stoniness and stiffness and excavation may prove expensive. However, it can be "expected to form a strong foundation bed

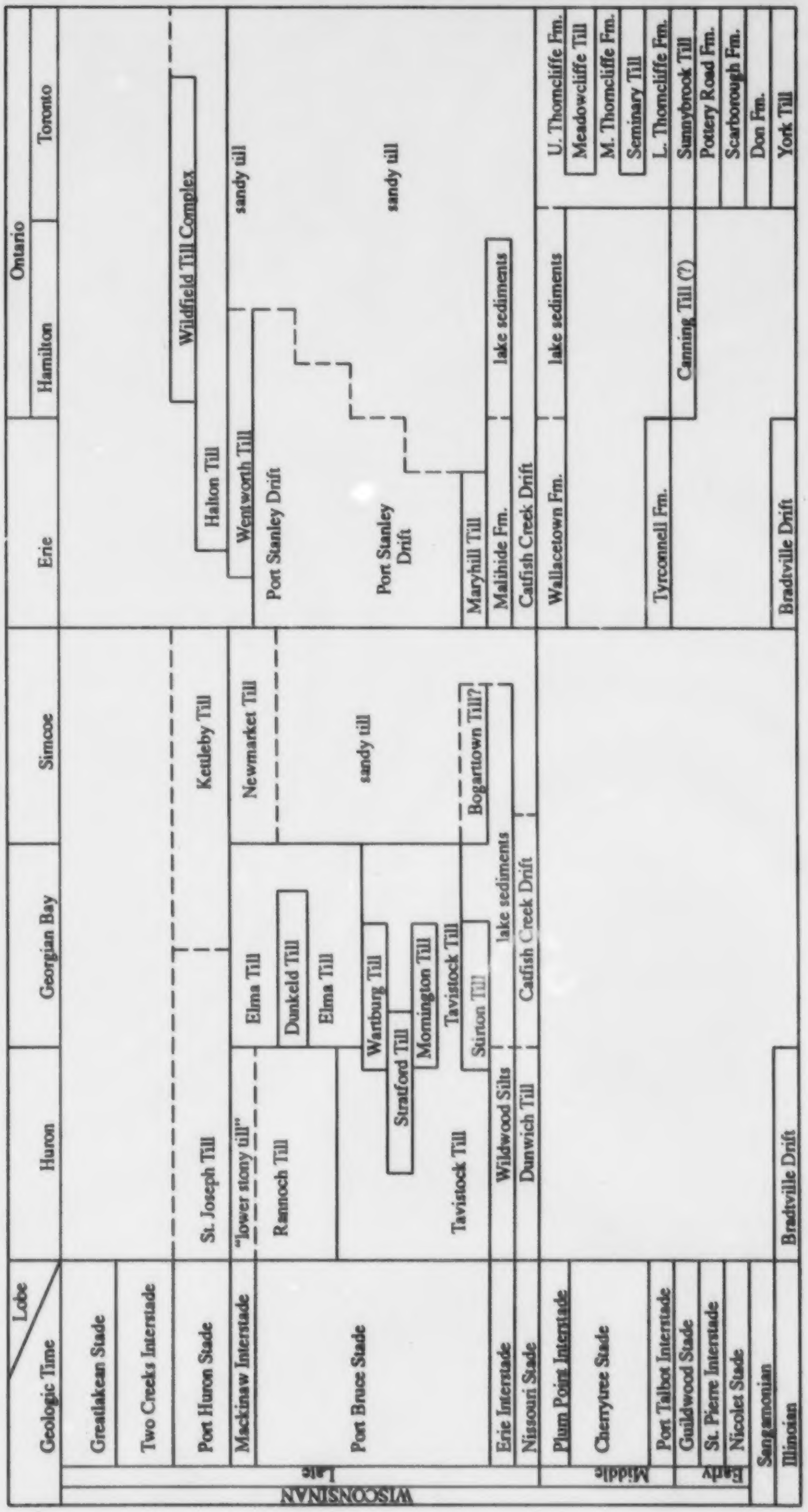

Geologic Time
Lobe
Huron
Georgian Bay
Simcoe
Erie
Ontario
Hamilton
Toronto
WISCONSINAN
Late
Middle
Early
Greatlakean Stade
Two Creeks Interstade
Port Huron Stade
Mackinaw Interstade
Port Bruce Stade
Erie Interstade
Nissouri Stade
Plum Point Interstade
Cherrytree Stade
Port Talbot Interstade
Guildwood Stade
St. Pierre Interstade
Nicolet Stade
Sangamonian
Illinoian
St. Joseph Till
"lower stony till"
Rannoch Till
Stratford Till
Tavistock Till
Wildwood Silts
Dunwich Till
Bradtville Drift
Elma Till
Dunkeld Till
Elma Till
Wartburg Till
Mornington Till
Tavistock Till
Stirton Till
lake sediments
Catfish Creek Drift
Kettleby Till
Newmarket Till
sandy till
Bogarttown Till?
Halton Till
Wentworth Till
Port Stanley Drift
Port Stanley Drift
Maryhill Till
Malahide Fm.
Catfish Creek Drift
Wallacetown Fm.
Tyrconnell Fm.
Bradtville Drift
Wildfield Till Complex
sandy till
sandy till
lake sediments
lake sediments
Canning Till (?)
U. Thorncliffe Fm.
Meadowcliffe Till
M. Thorncliffe Fm.
Seminary Till
L. Thorncliffe Fm.
Sunnybrook Till
Pottery Road Fm.
Scarborough Fm.
Don Fm.
York Till

Figure 21.35. Correlation chart for southwestern Ontario (*modified from Karrow 1989*).

either directly for foundations or as a base for piles" (Karrow 1988, p.190).

Indications from the Woodstock area (Cowan 1975) are that the initial movement of the ice sheet during the Nissouri Stade was controlled by the orientation of the Great Lakes basins, but changed to a more regional flow, from the north-northeast during the maximum advance of this stade. During ice recession as the ice sheet thinned, flow once again was controlled by the Great Lakes basins (Dreimanis 1961, 1964; Cowan 1975).

The Dunwich Drift, which was once considered to be Middle Wisconsinan, is now thought to be a Huron lobe facies of Catfish Creek Drift (Dreimanis and Barnett 1985). It is the oldest recognized member of the Catfish Creek Drift. Pre–Catfish Creek tills in the Kitchener area may represent slightly earlier fluctuations during the Nissouri Stade, but they may also be considerably older (Cooper 1975; Karrow 1989). The interlobate Sparta moraine may have formed in part during the initial advance of the lobes during the beginning of the Nissouri Stade. The Dorchester, Easthope, Waterloo and Orangeville moraines may have in part been formed during the breakup and lobation of Laurentide Ice Sheet during the latter part of the Nissouri Stade (Figure 21.36). Buried sand and gravel deposits at Leamington may also be related to this breakup.

## Erie Interstadial Sediments

The Erie Interstade is a period of ice-marginal recession along the southern edge of the Laurentide Ice Sheet in the lakes Michigan, Huron and Erie basins. A series of large

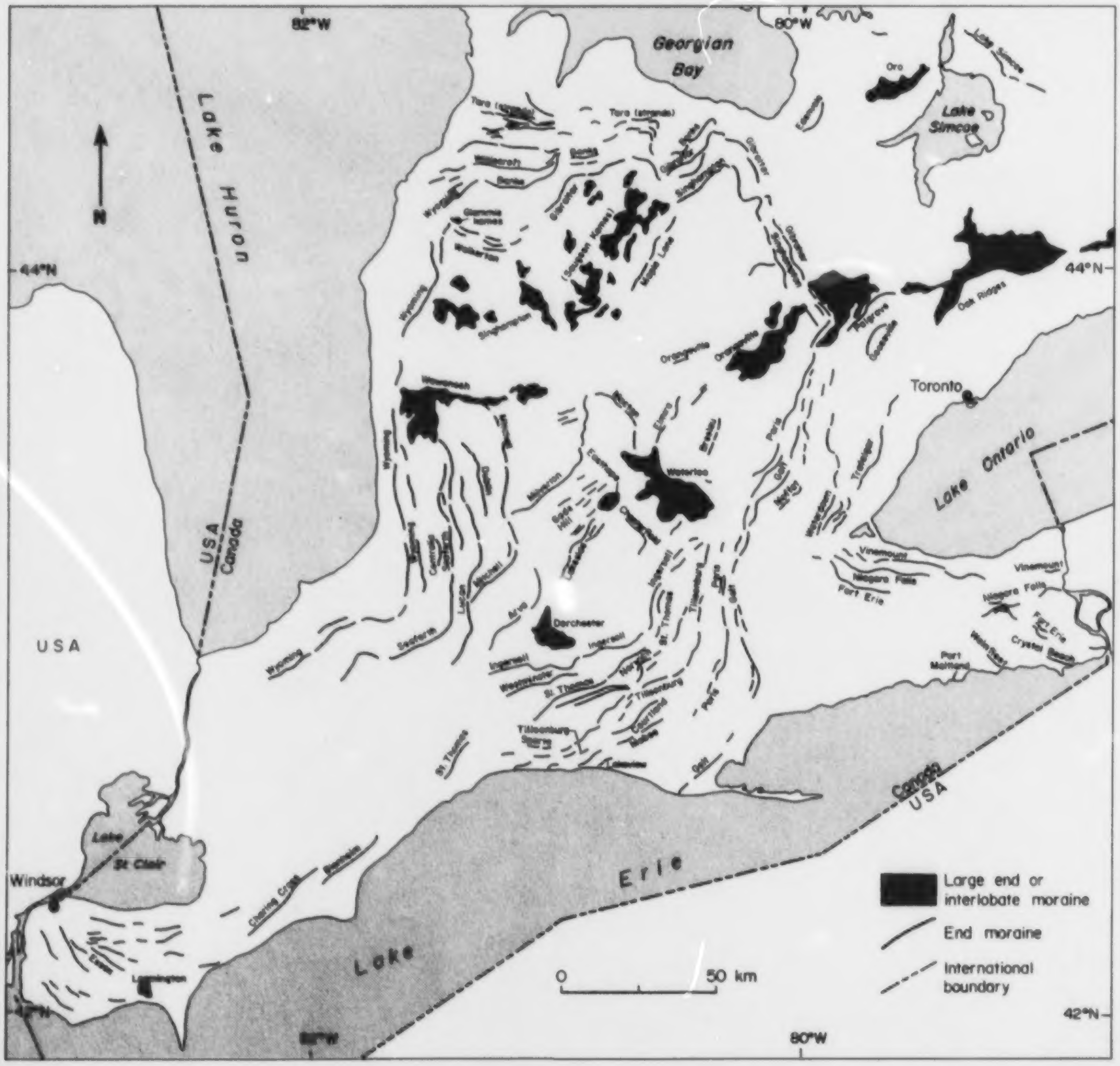

82°W
80°W
Georgian Bay
Oro
Lake Simcoe
N
Lake Huron
44°N
44°N
Tara
Wyoming
Singhampton
Gibraltar
Oak Ridges
Orangeville
Waterloo
Paris
Galt
Easthope
Milverton
Seaforth
Lucan
Mitchell
Dorchester
Ingersoll
St. Thomas
Tillsonburg
Toronto
Lake Ontario
Vinemount
Niagara Falls
Fort Erie
Crystal Beach
Port Maitland
USA
Canada
USA
Charing Cross
Lake St Clair
Windsor
Leamington
Lake Erie
Canada
USA
Large end or interlobate moraine
End moraine
International boundary
0
50 km
42°N
42°N
82°W
80°W

Figure 21.36. Major moraines of southwestern Ontario (many sources).

ice-contact proglacial lakes probably formed in the southern ends of these lake basins during this recession (Dreimanis 1969; Mörner and Dreimanis 1973). The record of these lakes is inadequate for detailed reconstruction of paleolake levels. Most of the record was destroyed by subsequent ice advances during the Port Bruce Stade. The fine-grained nature of the oldest tills deposited during the Port Bruce Stade indirectly confirms the existence of these lakes. Extensive incorporation of glaciolacustrine clay and silt is necessary to produce tills of such a fine-grained nature (Chapman and Putnam 1951; Dreimanis 1960; Barnett 1987).

Sand and gravel beach sediments were reported beneath Port Bruce Stadial till along the Lake Erie bluffs near Port Talbot (Mörner and Dreimanis 1973). These beach sediments were related to a lake (glacial Lake Leverett), whose level was similar to today's level of Lake Erie (Mörner and Dreimanis 1973). Mörner and Dreimanis (1973) suggested that the ice margin must have receded into the Lake Ontario basin during the Erie Interstade. The lack of fine-grained Port Bruce Stadial tills in the Owen Sound area led Sharpe and Edwards (1979) to suggest that the ice margin of the Huron and Georgian Bay lobes did not recede as far north as Chesley during the Erie Interstade (Figure 21.37).

A thick sequence of fine-grained glaciolacustrine rhythmites beneath till in the central part of the Lake Erie basin (Barnett 1987), and pollen-bearing silts termed the "Wildwood Silts" (Sigleo and Karrow 1977) that outcrop near St. Marys, have been assigned to the Erie Interstade. The Wildwood Silts, with a typical interstadial pine-spruce pollen assemblage, are thought to have been deposited in a local lake (Karrow 1984). Erie Interstadial deposits have yet to be dated, but are believed to be between 16 500 and 15 500 years old.

## Port Bruce Stadial Sediments

During the Port Bruce Stade, the Laurentide Ice Sheet once again advanced across the southern part of Ontario and extended well into the United States. The direction of ice flow was controlled by the orientation of the Great Lakes basins, with flow occurring radially outward within each basin. The Erie Interstadial glaciolacustrine sediments were overridden. The incorporation into the base of the glacier or subsole deformation of these fine-grained sediments must have been extensive. This is evidenced by the fine-grained nature of the Port Stanley Till deposited within the Lake Erie basin (Barnett 1987). The Tavistock Till, a Huron–Georgian Bay lobe till, is also very fine-grained south of Lake Huron and Lake St. Clair (R.I. Kelly and T. Morris, Ontario Geological Survey, personal communications, 1990).

The ice lobes coalesced and advanced to form the Union City–Powell moraine in the United States (Michelson et al. 1983). The northern extent of Erie–Ontario lobe tills on the surface is generally marked by the Ingersoll, Waterloo and Orangeville moraines, and the southern extent of Huron–Georgian Bay lobe tills is marked by the Blenheim, Ingersoll, Waterloo and Orangeville moraines. Several of these moraines, such as the Waterloo moraine, were constructed during both the Nissouri and Port Bruce stades (*see* Figure 21.36).

The Maryhill Till (Karrow 1974; map unit 4) and the Port Stanley Drift (de Vries and Dreimanis 1960) represent the fluctuations of the Erie–Ontario lobe during the Port Bruce Stade.

The Maryhill Till underlies Port Stanley Drift in the area between Woodstock and Kitchener and may occur in the Orangeville area (Cowan 1976). It is a strongly calcareous, silty clay to clay till of low to moderate plasticity. The Maryhill Till is massive and nonsorted, often exhibiting blocky structure. Clast content is low, usually less than 2%, and consists primarily of carbonate rock fragments. Matrix carbonate content averages approximately 35%. It occurs as ground moraine in the subsurface within the interlobate zone, forms the core of the Breslau moraine and outcrops along the southeastern flanks of the Waterloo moraine (Karrow, in press).

Port Stanley Drift consists of up to 5 layers of subglacially deposited Port Stanley Till (map unit 9) separated by glaciolacustrine sediments along the north-central shore of Lake Erie (Barnett 1982, 1987).

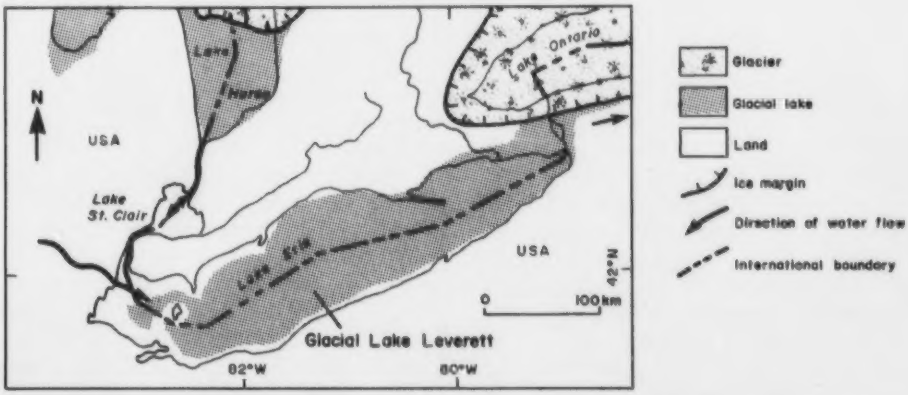


**Figure 21.37.** Estimate of ice-marginal retreat during the Erie Interstade (*modified from* Mörner and Dreimanis 1973).

Farther inland it is represented by only 1 layer of Port Stanley Till and glaciofluvial sediments. Laterally, the properties of Port Stanley Till vary greatly. Two main facies have been recognized; a southern fine-grained facies and a northern coarse-grained facies.

The northern facies of Port Stanley Till is a strongly calcareous, silt to sandy silt till derived from the overriding and incorporation of pre-existing glaciofluvial sand and gravel deposits and dolostone bedrock. It is nonplastic or has a low plasticity. Matrix carbonate content averages 45% and is strongly dolomitic. Clast content ranges between 10 and 30% and dolostone clasts are dominant. The northern facies of Port Stanley Till is commonly associated with glaciofluvial outwash sediments. It usually displays fissility and may contain sandy lenses or inclusions. This facies of the Port Stanley Till occurs as ground moraine or in drumlins of the Guelph drumlin field. Its thickness varies from 2 to 15 m.

The southern facies of Port Stanley Till is a strongly calcareous, clayey silt to silty clay till with low plasticity. This till is usually massive with blocky conchoidal fracture. In places, it is vertically jointed. Matrix carbonate content averages 30 to 40% and calcite is predominant. Clast content is less than 2%; clasts are predominantly dolostone where the till overlies Silurian bedrock and limestone where the till overlies Devonian bedrock.

The southern facies of Port Stanley Till is commonly interbedded with glaciolacustrine sediments and can be associated with thinly bedded flowtills in ice-marginal settings. It occurs as ground moraine and in end moraines with thicknesses of till not exceeding 25 m (usually between 2 and 10 m thick).

During ice-marginal recession of the Erie–Ontario lobe toward the southeast into the Lake Erie and Ontario basins, a series of end moraines were formed that mark positions of standstill or minor readvances of the ice margin. This series of moraines includes the Ingersoll, Westminster, St. Thomas, Norwich, Tillsonburg, Courtland, Mabee and Lakeview moraines (*see* Figure 21.36). The southernmost parts of these moraines formed within large ice-contact glacier-fed lakes.

Tills deposited during the Port Bruce Stade by the combined Huron–Georgian Bay lobe include the Stirton, Tavistock, Mornington, Stratford and Wartburg tills (Karrow 1974, 1989). The Huron and Georgian Bay lobes separated during the later part of this stade and acted independently, depositing the Elma Till (Georgian Bay lobe) and the Rannoch Till (Huron lobe).

The Stirton Till (Karrow 1974; not shown on accompanying Map 2556) is a strongly calcareous, very stiff, clayey silt to silty clay till with low to moderate plasticity. Matrix carbonate content averages between 37 and 47%. Stirton Till is predominantly dolomitic where it overlies Silurian bedrock and is calcareous where it overlies Devonian limestones. Clast content is low, usually less than 1%. Carbonate clasts are by far the most abundant. Clast fabric determinations and the orientation of striations on lodged cobbles indicate a direction of ice flow during deposition from the northwest to the southeast.

Stirton Till occurs primarily in the subsurface within the Conestogo River valley as a layer of ground moraine 1 to 3 m thick, usually overlain by Tavistock Till. The Stirton Till is probably the product of a surge or enhanced flow rates, based on its limited distribution (Cowan 1979). This till unit is considered by some to be the basal member of the Tavistock Till (Cowan 1979).

The properties and characteristics of the Tavistock Till (Karrow 1974; map unit 5) change greatly along its outcrop length. It is a highly calcareous, silty clay to clayey silt till of low to medium plasticity in the area south of Lake Huron and Lake St. Clair. Clast content is low, usually less than 5% and is dominated by carbonate and shale clasts. Clasts of Huronian tillite and jasper conglomerate are common locally. This facies of the till occurs as ground moraine, 2 to 15 m thick, commonly interbedded with glaciolacustrine sediments.

Farther north in the London and Woodstock area, the Tavistock Till is a strongly calcareous, silt to sandy silt to sand till. It is nonplastic or has low plasticity. Commonly having a fissile structure, this till unit may contain local intercalations of stratified sediments. Matrix carbonate content averages 41%. Clast content ranges from 5 to 10% with limestone clasts dominant. Huronian tillite and jasper conglomerate erratic clasts are common locally.

The Tavistock Till is coarser grained in this area because it is derived from underlying glaciofluvial sediments or Catfish Creek Till (Cowan 1975). It occurs as ground moraine or in drumlins and rarely exceeds 12 m in thickness. Tavistock Till is overlain by glaciofluvial or glaciolacustrine sediments or younger tills.

Tavistock Till found north of Waterloo is a strongly calcareous silty clay to silt till largely derived from glaciolacustrine sediments. It has massive to blocky structure, is very stiff and has low plasticity. Matrix carbonate averages 35 to 45% with dolomite predominating in the north and calcite in the south. Clasts are dominantly carbonates and make up about 2% of the till. This northern facies of Tavistock Till is between 2 and 12 m thick and occurs as ground moraine which may be fluted or drumlinized. It is the surface sediment along the western flank of the Orangeville moraine. Other moraines formed during the retreat of the Huron–Georgian Bay lobe composed in part of Tavistock Till include the Blenheim, Waterloo and the Arva moraines (*see* Figure 21.36).

The Mornington Till (Karrow 1974; map unit 6) is a strongly calcareous, silty clay till derived from underlying glaciolacustrine silt and clay and older tills. It has low to medium plasticity and is nonsorted and nonstratified. Matrix carbonate content averages 35 to 40%, the majority of which is calcitic. Dolomite content does increase northward, however. The Mornington Till is clast poor and the main lithologies present are carbonates. The Mornington Till occurs as flat and weakly fluted ground moraine varying between 1 and 3 m in thickness. This till is believed to form the core of the Macton moraine and is considered by some to be an upper member of Tavistock Till (Cowan 1979).

The Stratford Till (Karrow 1974; map unit 7) overlies the Tavistock Till. It is a strongly calcareous, sandy silt to silt till with low plasticity. The Stratford Till is soft, nonsorted and nonstratified with little structure. Matrix carbonate content averages 50% with about equal amounts of calcite and dolomite. Limestone is the dominant clast lithology, in a till which contains about 5 to 10% clasts. The Stratford Till occurs as a thin ground moraine sheet, 1 to 3 m thick, and in a series of *en échelon*, small, linear ridges called the "Gads Hill moraines" (Karrow, in press). The Stratford Till is commonly overlain by thin deposits of glaciolacustrine silt and clay.

The Wartburg Till (Karrow 1974; map unit 8) is a strongly calcareous, silty clay till with low to moderate plasticity. It is very poorly sorted and nonstratified. Matrix carbonate content averages 40 to 50% with calcite predominating. Clast content is less than 1% with limestone being the dominant rock type represented. The stratigraphic relationships of the Wartburg Till are poorly known. It forms the core of the Milverton moraine and occurs as ground moraine that is largely buried by Elma Till. Thicknesses of this till sheet range from 2 to 10 m.

The Elma Till (Karrow 1974; map unit 10) is a deposit of the Georgian Bay lobe. It is a strongly calcareous silt, sandy silt and clayey silt till. It is generally more clayey in the south and more sandy in the north. The Elma Till is nonplastic or has a low plasticity. It often exhibits a fissile structure. Matrix carbonates average 45 to 55%; the till is strongly dolomitic in the north, while calcite predominates in the south. Clast content ranges from 5 to 25%. Dolomite clasts are abundant in samples of till from the north; however, limestone predominates in southern till samples.

The Elma Till occurs as ground moraine and in drumlins of the Teeswater drumlin field. It is associated with the Singhampton moraine (formerly the Saugeen Kames) and is overlain by glaciofluvial sand and gravel, glaciolacustrine silts and younger tills. Elma Till ranges in thickness between 2 and 15 m. It was deposited during the latter part of the Port Bruce Stade, but deposition of this till probably continued during the following Mackinaw Interstade.

The Rannoch Till (Karrow 1977; map unit 11), a Huron lobe till, is a strongly calcareous, silt to silty clay till with low plasticity. This till tends to become finer grained westward. It is nonsorted and nonstratified, commonly with a blocky structure. Matrix carbonate content averages 50 to 60% with calcite predominating in the south and dolomite in the north. Clast content is usually less than 2% with locally derived limestone making up 85 to 95% of the clasts in the south and 60% in the north. Pebble fabrics, striated tops of lodged boulders and surface flutings indicate radial flow out of the Lake Huron basin.

The Rannoch Till occurs as ground moraine and in several end moraines including the Mitchell, Dublin, Lucan, Seaforth and Centralia moraines (*see* Figure 21.36). It can be up to 70 m thick, but is typically from 2 to 6 m thick and covered by glaciolacustrine silt and sand or local supraglacial (ablation) tills.

South of Lake Huron, a calcareous silty clay to clay till containing numerous lenses and inclusions of glaciolacustrine sediments occurs (informally referred to as the southern till, Cooper 1979). This till is nearly stone-free with limestone clasts being the most abundant. It is considered by some to be an upper member of the Rannoch Till, and is found as ground moraine and as a veneer on the southern part of the Seaforth moraine (Cooper 1979). On the accompanying map (*see* Map 2556, map case), this till is included with the Rannoch Till. Shoreline deposits of glacial Lake Arkona overlie this till in the Arkona area.

The Dunkeld Till (Cowan and Pinch 1986; map unit 12) is a strongly calcareous silt till. Matrix carbonate content averages 48%; dolomite is the dominant carbonate mineral. Dolostone clasts are the most abundant in this till. Dunkeld Till occurs as ground moraine within the Saugeen River valley and is in the core of the Walkerton moraine. Dunkeld Till is the product of a minor readvance of the ice margin over glaciolacustrine silts of glacial Lake Saugeen. Elma Till is probably both older and younger than the Dunkeld Till.

The Newmarket Till (Gwyn 1972; map unit 13) is a till deposited by the Simcoe lobe during the latter part of the Port Bruce Stade. It is younger than the Port Stanley Till and older than the Wentworth Till in its marginal area, based on cross-cutting relationships of end moraines in the Orangeville area (Cowan 1976).

The Newmarket Till is a calcareous silt to sandy silt till. It is nonplastic and ranges from soft to very stiff. Matrix carbonate content averages 30 to 40% with calcite predominant east of the Niagara Escarpment and dolomite dominant above the escarpment. Clast content varies from 10 to 45%. The Newmarket Till occurs as hummocky end moraine, as in the Singhampton moraine, or as ground moraine, which is often covered by glaciofluvial or glaciolacustrine sands and silts. This till is usually less than 3 m thick, but ranges up to 12 m in thickness.

During ice-marginal recession, meltwater was ponded in the western end of the Lake Erie basin and in the southern end of the Lake Huron basin between the Mississippi River–Great Lakes drainage divide and the receding ice margin forming glacial Lake Maumee I. Water levels rose within these basins until the water began to spill down the Wabash River in Indiana into the Mississippi River system (Eschman and Karrow 1985). During the life of this outlet, glacial Lake Maumee I existed in front of the receding ice lobes' margins. The glacial Lake Maumee II stage began with the farther northward retreat of the Huron lobe margin (Leverett and Taylor 1915). The Imlay channel was uncovered, allowing meltwater to flow across the thumb of Michigan into Saginaw Bay drainage basin, then down the Grand River into a high level lake in the Lake Michigan drainage basin. The level of the lake in the Lake Michigan drainage basin was controlled by the Des Plaines River, sometimes referred to as the Chicago Outlet.

A readvance to the St. Thomas moraine (Port Stanley Till) by the Erie lobe and possibly to the Mitchell moraine (Rannoch Till) may have closed the Imlay channel, raised the water level and diverted meltwater outflow down the Wabash River again (glacial Lake Maumee III, Figure 21.38). During the subsequent ice-marginal retreat, the Imlay channel was reoccupied (glacial Lake Maumee IV) and progressively lower outlets across and around the thumb of Michigan opened. An early Lake Whittlesey (Ubly Channel; L.J. Chapman, Thornbury, personal communication, 1989) and several levels of glacial Lake Arkona (Grand River) formed (Eschman and Karrow 1985).

The Komoka Delta at London was being built during the Maumee lake levels from sediments carried by meltwater streams that flowed down the valleys now occupied by the Thames River and some of its tributaries. The delta at Coldstream near London probably formed in glacial Lake Maumee IV as the Huron lobe was forming the Seaforth moraine. The ice margin of the Erie lobe was probably forming the Mabee and Lakeview moraines during glacial Lake Maumee IV (Figure 21.39).

Thick rhythmite sequences of clay, silt and sand were deposited within the paleolake basins and underlie some of the extensive, flat, lake plains in southwestern Ontario (map units 24 and 25). Shoreline deposits of the various levels of glacial Lake Maumee have been identified between London and Delhi (Barnett 1985). Several interesting summaries of the postglacial lake sequences in the Great Lakes basins have been written which provide additional ideas (Leverett and Taylor 1915; Hough 1958; Prest 1970; Fullerton 1980; Karrow and Calkin 1985).

Patterned ground, evidence of permafrost conditions, covered parts of the deglaciated landscape above the level of the postglacial lakes (Morgan 1972, 1982).

## Mackinaw Interstadial Sediments

With the continuation of the ice-marginal retreat of the Laurentide Ice Sheet, a regional lowering of lake level in the lakes Huron and Erie basins occurred. This period of general ice retreat and possible climatic warming is referred to as the Mackinaw Interstade (Dreimanis and Karrow 1972). Glacial Lake Arkona formed in the combined lakes Huron and Erie basins. Shorelines of glacial Lake Arkona, of which there are 3 recognized levels in the United States, are well developed on Port Bruce Stadial tills such as the Rannoch Till (defined to include the southern till; Cooper 1979) of the Huron lobe and the Port Stanley Till of the Erie–Ontario lobe.

The relationship of glacial Lake Arkona deposits to the Paris moraine, which is composed of Wentworth drift, is unclear. It appears that this moraine was either forming in glacial Lake Arkona or a minor advance to this moraine overrode Arkona sediments. Lake levels had fallen substantially, however, by the time the Galt moraine (Wentworth drift) had formed (Barnett 1985). The Paris and Galt moraines, composed of Wentworth drift, are believed

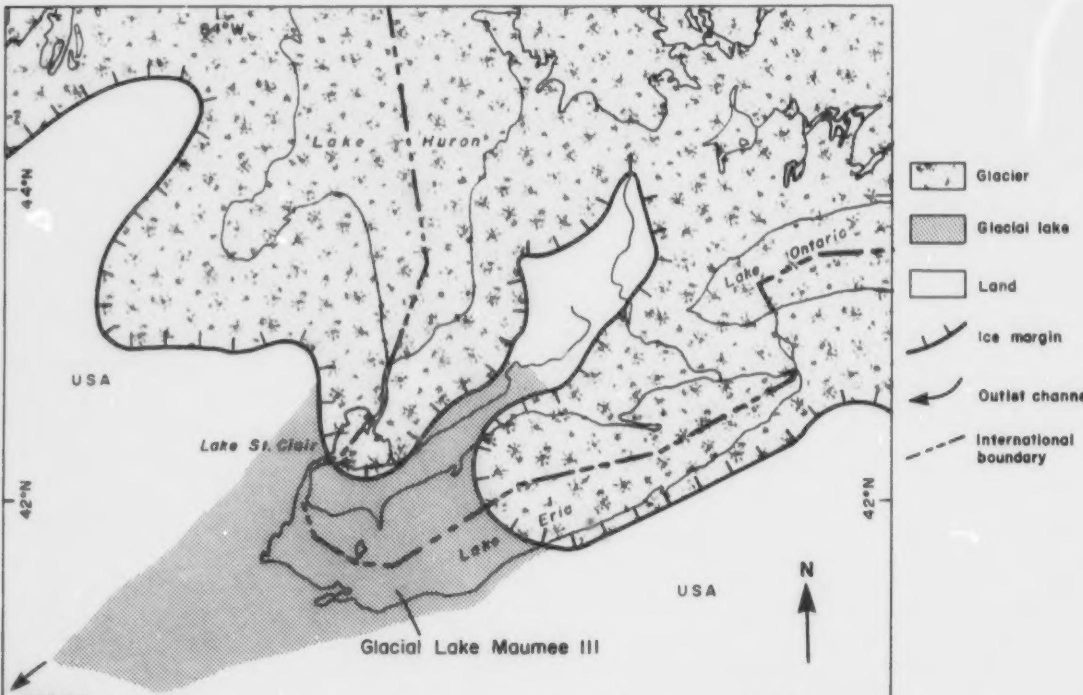


Figure 21.38. Glacial Lake Maumee III.

to be the result of a period of readjustment of the Erie–Ontario lobe margin during the Mackinaw Interstade. Part of the Elma and Newmarket till sheets were probably deposited during this interstade as well.

The Wentworth Till (Karrow 1959; redefined *in* Karrow 1974; map unit 14) is a strongly calcareous, sandy silt till in the north where glaciofluvial sediments, sandy tills or carbonate rocks were incorporated, and a clayey silt to silty clay till in the south along the Lake Erie shoreline where fine-grained glaciolacustrine sediments were incorporated. It ranges from being nonplastic to having a low plasticity. The matrix contains 30 to 40% carbonates, which are predominantly dolomite over Silurian bedrock, and calcite over Devonian bedrock. Clast content averages 7 to 10%; this varies from less than 2% when the till is fine textured and up to 30% when the till is very sandy. The Wentworth Till is associated with outwash gravel and sand in the north and with glaciolacustrine sandy fossil-bearing rhythmites in the Lake Erie basin. The Wentworth Till occurs as ground moraine, in drumlins of the Caledonia and Flamborough drumlin fields, and in the Paris, Galt and Moffat moraines.

During the formation of the Galt moraine, about 13 400 years ago, water levels had fallen to the Jacksonburg delta level (Barnett 1985). Eastern outlets remained blocked by the Erie–Ontario lobe. Drainage must have been westward across the Indian River lowlands in Michigan or via the straits of Mackinac. Most of the state of Michigan was ice-free at this time (Miller and Benninghoff 1969).

With farther retreat of the ice margin northward, eventually into the Lake Ontario basin, eastern outlets were uncovered which created lower and lower lake levels in the lakes Huron and Erie basins (Lake Ypsilanti, Kunkle 1963; Figure 21.40). North of Lake Ontario, the Ontario lobe was separating from the Simcoe lobe and glaciofluvial and glaciolacustrine sediments were beginning to be deposited into this interlobate zone (Oak Ridges moraine).

The large valleys in the area near Barrie and east of Lake Simcoe to Trenton are believed to be tunnel valleys. This network of tunnel valleys was cut by a catastrophic release of subglacially stored meltwater (jökulhlaup), possibly during the Mackinaw Interstade. The material eroded during the formation of the tunnel valleys may have been deposited in the Oak Ridges moraine and in the glaciofluvial outwash systems in front of or west of the Paris and Galt moraines and south of the Singhampton and Gibraltar moraines. The tunnel valleys appear to terminate at the heads of these outwash systems and on the north side of the Oak Ridges moraine (Barnett 1990). The tunnel valleys in the Trenton area, in places, appear to cross the Oak Ridges

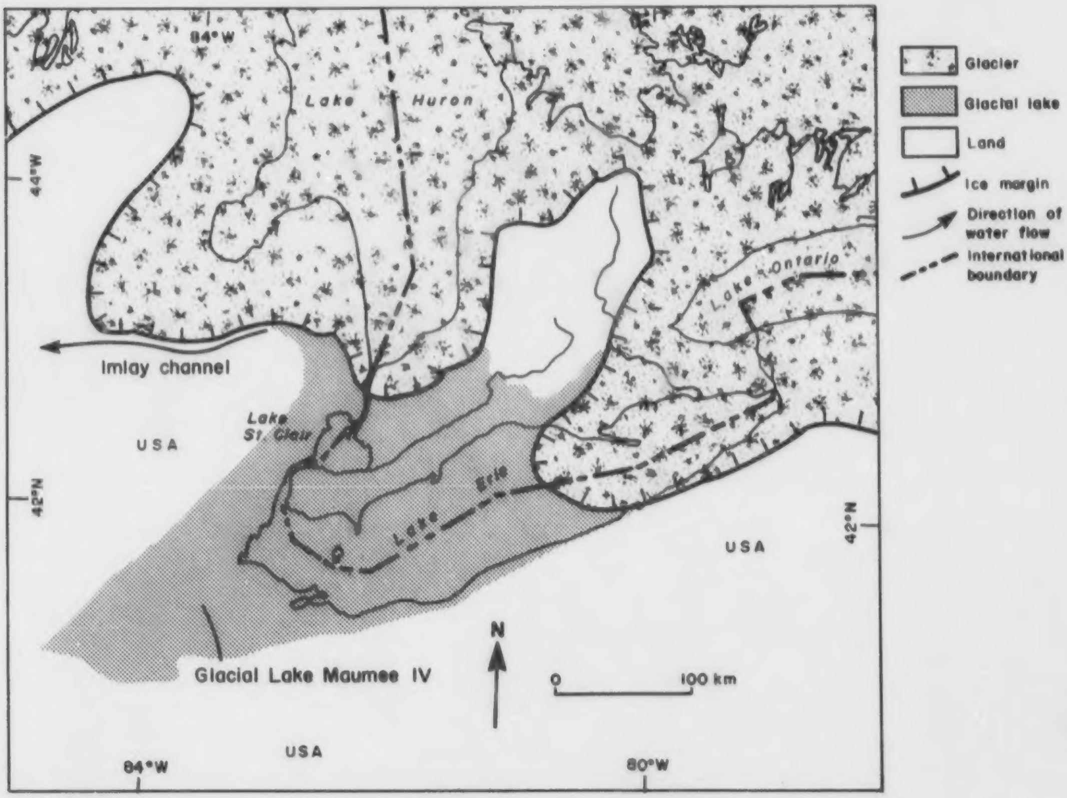


Figure 21.39. Glacial Lake Maumee IV.

moraine and may be slightly younger at the east end of the moraine (G.A. Gorrell, Gorrell Resource Investigations, personal communications, 1990).

During the Mackinaw Interstade, the deglaciated part of southern Ontario (Ontario Island) supported a wide variety of plants with modern arctic, boreal and prairie affinities. Initial populations of spruce had been established (Terasmae 1981; Warner and Barnett 1986). Permafrost conditions, which were widespread during the Port Bruce Stade, apparently ameliorated during the Mackinaw Interstade (Karrow 1989).

## Port Huron Stadial Sediments

During the Port Huron Stade, the Laurentide Ice Sheet margin readvanced into the eastern end of the Lake Erie basin and to the southern end of Lake Huron. The Halton Till (Erie–Ontario lobe), the Kettleby Till (Simcoe lobe) and the St. Joseph Till (Huron and Georgian Bay lobes) were deposited during this advance. The Wyoming moraine, Banks moraine and Waterdown moraines roughly delineate the extent of this advance, which closed the eastern outlets and raised water levels in the lakes Huron and Erie basins to that of glacial Lake Whittlesey (Ubly Channel outlet, Michigan), 13 000 years ago (Figure 21.41).

The St. Joseph Till (Cooper and Clue 1974; map unit 15) is a strongly calcareous silt to silty clay till of low plasticity. It tends to become finer grained southward. Extensive incorporation and subsole deformation of glaciolacustrine sediments deposited during the Port Bruce Stade and Mackinaw Interstade has occurred to produce this fine-grained till. Blocky structure and columnar jointing are common in exposures of the St. Joseph Till. Matrix carbonate content averages 45%. Clast content is low, between 1 and 2%. Key erratics are tillite and jasper conglomerate, which have been transported from the area north of Lake Huron. The St. Joseph Till of the Georgian Bay lobe exhibits minor textural differences with respect to the St. Joseph Till of the Lake Huron basin. Mineralogical variations are a result of source area differences.

The St. Joseph Till occurs in the Wyoming moraine which parallels the shore of Lake Huron, in the Banks and Williscroft moraines of the Georgian Bay lobe, and as associated ground moraine. It can be overlain by outwash sand and gravel and glaciolacustrine gravel, sand and silt.

The Kettleby Till (Gwyn 1972; map unit 16) represents a southerly advance of the Simcoe lobe to the Oak Ridges moraine and Niagara Escarpment during the Port Huron Stade. The Kettleby Till is a highly calcareous silty clay to clay till. It is usually clast-poor, averaging less than 1% clasts. It often exhibits a blocky structure in exposures and has moderate plasticity. The Kettleby Till is usually less than 2 m thick and occurs as a discontinuous sheet of ground moraine overlying older topography. It is the surface till along the northern flank of the western end of the Oak Ridges moraine.

The Erie–Ontario lobe deposited the Halton Till (Karrow 1959; map unit 17) during the Port Huron advance and the Wildfield Till Complex (White 1975) during ice-marginal recession. The Halton Till is strongly calcareous and has a low plasticity. The till is texturally variable as follows: it is primarily a sandy silt to silt till where overlying rocks of the Georgian Bay Formation; and, a clayey silt to silty clay till south of the Niagara Escarpment

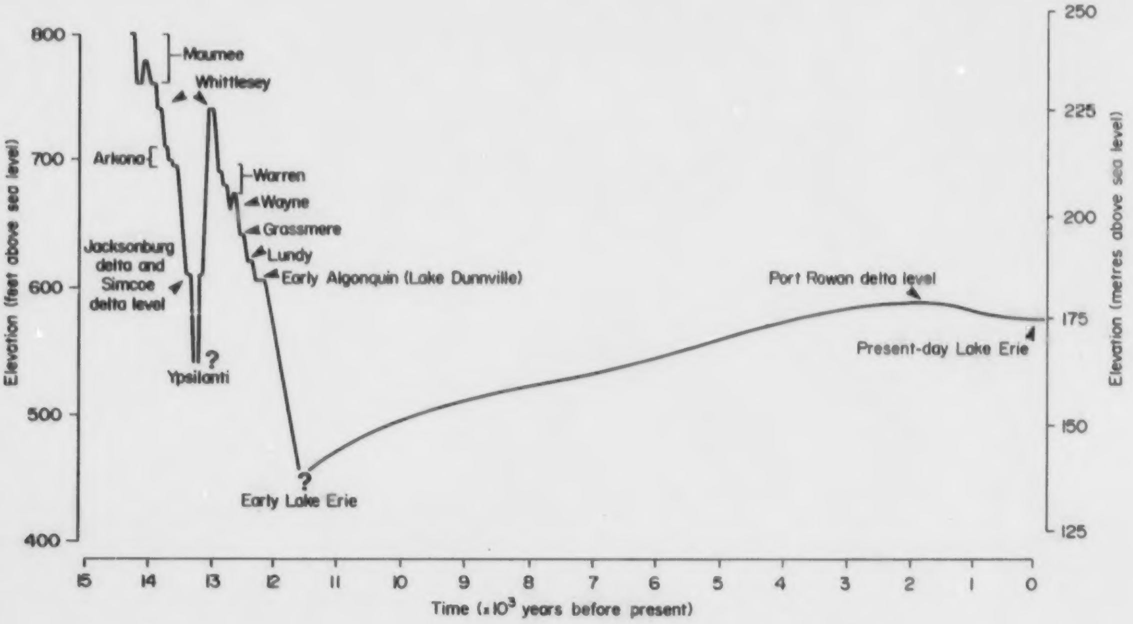


**Figure 21.40.** Ancestral lake phases in the Lake Erie basin (*modified from* Barnett 1985).

where Queenston Formation shale and glaciolacustrine silt and clay have been overridden. Clast content varies between 2 and 12%; higher near the Niagara Escarpment. Matrix carbonate ranges from 15 to 35% with calcite being predominant on Ordovician and Devonian bedrock terrains and dolomite on Silurian bedrock terrain.

The Halton Till occurs as low-relief ground moraine north of Hamilton and in a series of end and recessional moraines above the Niagara Escarpment. Within the Niagara Peninsula, the Halton Till is the core of the Port Maitland, Wainfleet, Crystal Beach, Fort Erie, Niagara Falls and Vinemount moraines. The Waterdown moraines and the Palgrave moraine north of Hamilton are cored by this till (*see* Figure 21.36). Halton Till discontinuously covers the southern flank of the western end of the Oak Ridges moraine.

Shoreline deposits and features of glacial Lake Whittlesey are well developed in Ontario and occur north of Lake Erie beyond the margin of the Halton Till (Barnett 1979). The Summit delta, located near Ancaster, formed in glacial Lake Whittlesey at the limit of the Erie–Ontario lobe advance during the Port Huron Stade. Large deltas were built into glacial Lake Whittlesey at Brantford and west of London by extensive braided stream systems flowing down the Grand and Thames river valley, respectively.

Falling lake levels in the lakes Huron and Erie basins accompanied ice-marginal recession from the Port Huron maximum (Hough 1958; Barnett 1985; Calkin and Feenstra 1985; Eschman and Karrow 1985). Glacial Lake Warren I is the first of 3 levels of Lake Warren that followed glacial Lake Whittlesey in the lakes Huron and Erie basins. It drained westward down the Grand River of Michigan. Downcutting of this outlet led to the development of glacial lakes Warren II and Warren III. The position of the ice margin in the Lake Huron basin is uncertain, but was behind or west of the Wyoming moraine in the southern part of the basin. The Erie–Ontario lobe, however, was probably standing at the Crystal Beach or Fort Erie moraine during Lake Warren I and the Niagara Falls moraine during Lake Warren II. The large ice-marginal fan-delta at Fonthill was constructed in a small re-entrant in the glacier margin during the Warren I and II stages (Feenstra 1981).

Glacial Lake Wayne is a lower water stage that is believed to have occurred between glacial lakes Warren II and Warren III. Glacial Lake Wayne drained eastward, probably through the Syracuse Channels as they became uncovered by the glacier. Readvance of the ice margin, at least to the Batavia Moraine in the state of New York (Calkin and Feenstra 1985) or possibly to the Vinemount moraine, raised the water level to the glacial Lake Warren III level. The Warren III level is marked by a well-defined shore bluff

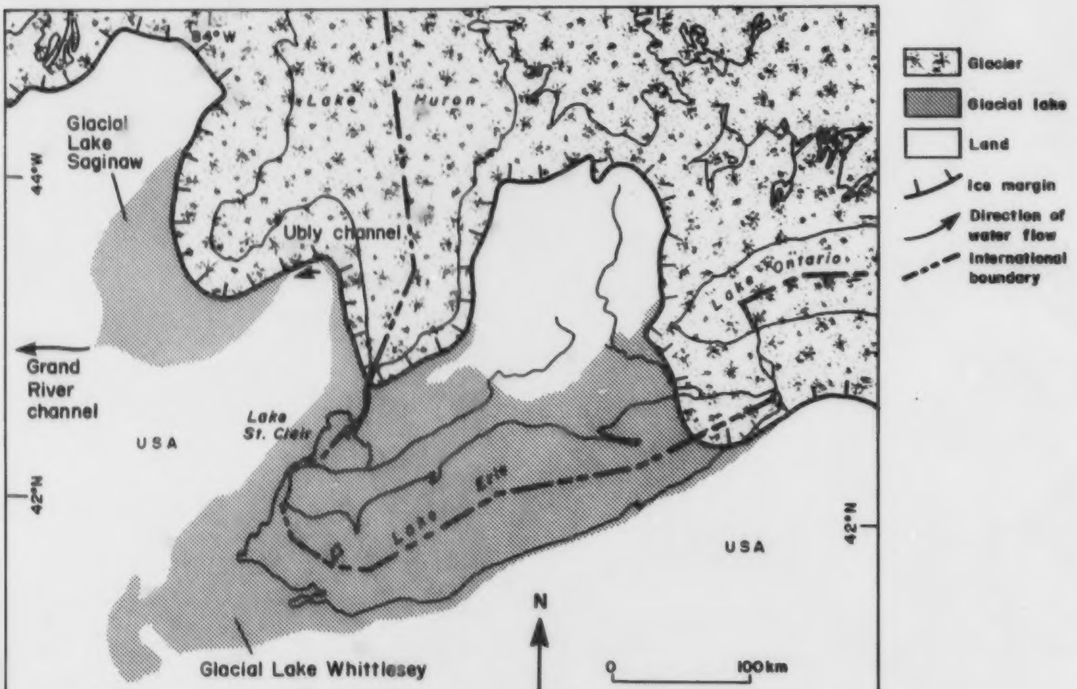


**Figure 21.41.** Glacial Lake Whittlesey (*modified from* Calkin and Feenstra 1985).

and wave-cut platform near the crest of Fonthill (Feenstra 1981). Well-developed glacial Lake Warren shoreline deposits and landforms occur at Ridgetown, Leamington and along the Wyoming moraine.

Additional ice-marginal retreat brought about the final abandonment of the Grand River valley as an outlet for the glacial Great Lakes. Drainage was probably eastward during the subsequent glacial Lake Grassmere, and then westward again for glacial Lake Lundy via the Straits of Mackinac (Figure 21.42; Burgis 1977). Shoreline deposits for these glacial lakes are generally weakly developed in Ontario; however, small abandoned deltas have formed along several of the creeks that flow into Lake Erie. Shoreline features occur near Leamington and around the south end of Lake St. Clair and Lake Huron (Eschman and Karrow 1985).

Thick deposits of glaciolacustrine sediments were deposited during the series of proglacial lakes described above (map units 24 and 25). The glaciolacustrine sediments generally coarsen upward with clay and silt rhythmites overlain by silts and sands. The large sand plains of Caradoc, Bothwell and Norfolk, and the clay plains of St. Clair, Ekfrid and Haldimand (Chapman and Putnam 1984) are underlain by glaciolacustrine sediments deposited during this series of falling lake levels in the lakes Erie and Huron basins.

Local ponding occurred between the northern margin of the Erie–Ontario lobe and the Oak Ridges moraine and the Niagara Escarpment (Peel ponding or glacial Lake Peel; *see* Figure 21.42), and between the southern margin of the ice lobe and the Niagara Escarpment (Bell Terraces, Feenstra 1981). A readvance of the ice margin north of Lake Ontario into glacial Lake Peel deposited the Wildfield Till Complex and formed the Trafalgar moraine.

The Wildfield Till or Wildfield Till Complex (White 1975) is a strongly calcareous silty clay to clay till of low to intermediate plasticity. Clast content is usually less than 2% and consists mainly of limestone, siltstone and shale clasts. Matrix carbonate content up to 35% has been reported, of which calcite is the dominant carbonate mineral. The

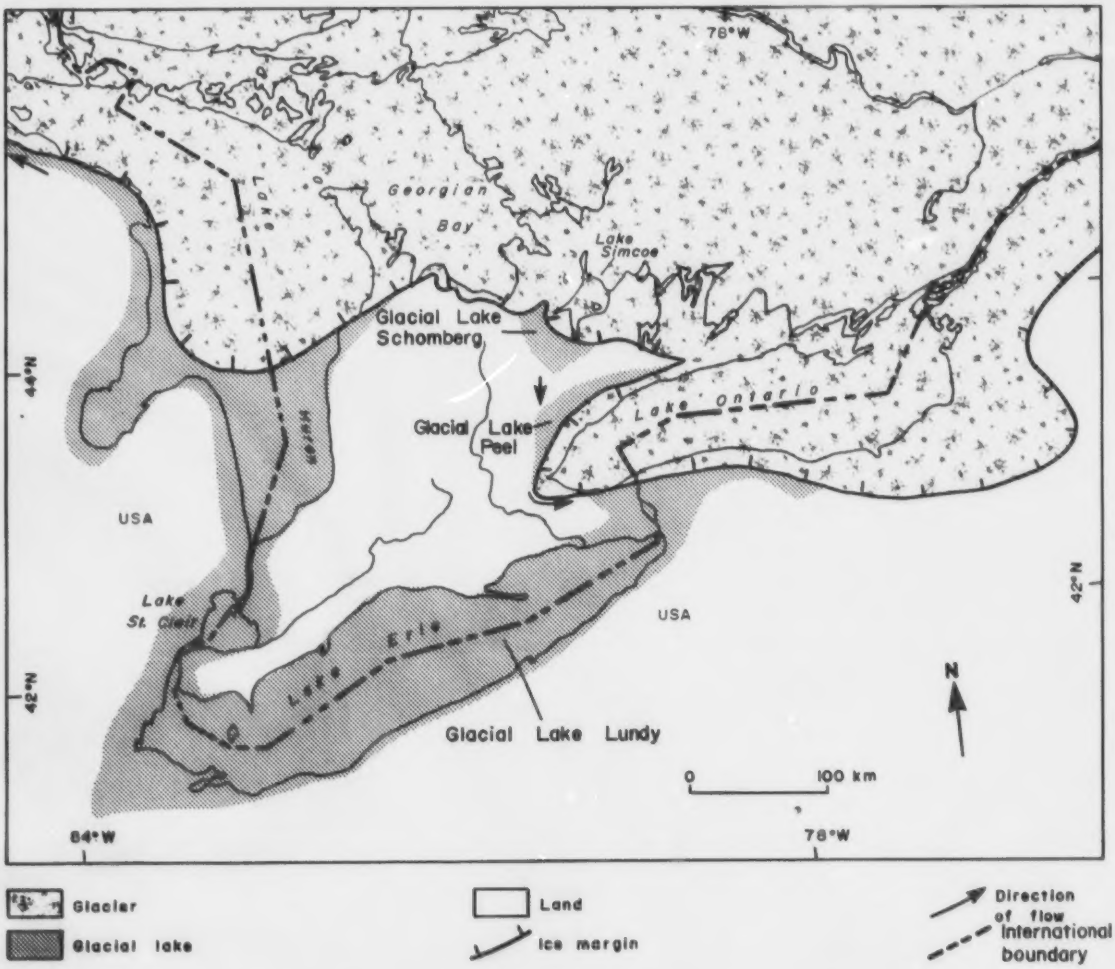

78°W
Lake Huron
Georgian Bay
Lake Simcoe
Glacial Lake Schomberg
Glacial Lake Peel
Lake Ontario
44°N
USA
Lake St. Clair
Lake Erie
USA
42°N
42°N
N
Glacial Lake Lundy
0
100 km
84°W
78°W
Glacier
Glacial lake
Land
Ice margin
Direction of flow
International boundary

**Figure 21.42.** Glacial lakes Lundy, Peel and Schomberg (*modified from* Chapman and Putnam 1984).

Wildfield Till is derived from glaciolacustrine sediments of glacial Lake Peel and commonly contains clasts of this material. This till is usually interbedded with glaciolacustrine sediments, leading some to suggest that it consists predominantly of flowtills (Sado et al. 1984). It occurs as low-relief ground moraine, often buried beneath younger glaciolacustrine silts and clays. Where exposed at the surface, the Wildfield Till has been included with the Halton Till on the accompanying map (map unit 17; *see* Map 2556, map case).

North of the Oak Ridges moraine, ponding also occurred between the moraine, the receding Simcoe lobe margin and the Niagara Escarpment (glacial Lake Schomberg; *see* Figure 21.42). Thick clay and silt rhythmites were deposited in glacial Lake Schomberg. A layer of Kettleby Till interfingers with glacial Lake Schomberg deposits along the northern flank of the Oak Ridges moraine.

## Two Creeks Interstadial Sediments

The Two Creeks Interstade is a general period of ice-marginal recession during which the margin of the Laurentide Ice Sheet continued to recede northward, about 12 000 years ago. Separate lakes formed in the Lake Ontario (glacial Lake Iroquois) and Lake Erie (Early Lake Erie) basins, and waters in the lakes Huron and Michigan basins coalesced (glacial Lake Algonquin, Figure 21.43). Minor oscillations of the ice margin resulted in the deposition of fine-grained tills by the Ontario lobe in the Trenton area and by the Georgian Bay lobe in the Edenvale area (map unit 21). Elsewhere, sandy silt to silt till (map unit 19) was deposited on areas underlain by Paleozoic carbonate rocks and sandy till (map unit 18) on the southern Canadian Shield (*see* General Characteristics of Ontario Tills section for a description of these 2 till types).

While the Dummer moraine was forming, major ice-marginal stagnation occurred along the Black River–Trenton escarpment that separates the Precambrian rocks of the Canadian Shield from the Ordovician rocks of the Great Lakes–St. Lawrence Lowlands. The till of the Dummer moraine (map unit 20) is remarkable for its stoniness. Numerous large angular blocks of limestone that were quarried by the glacier and carried only a short distance before being deposited within the hummocks of the

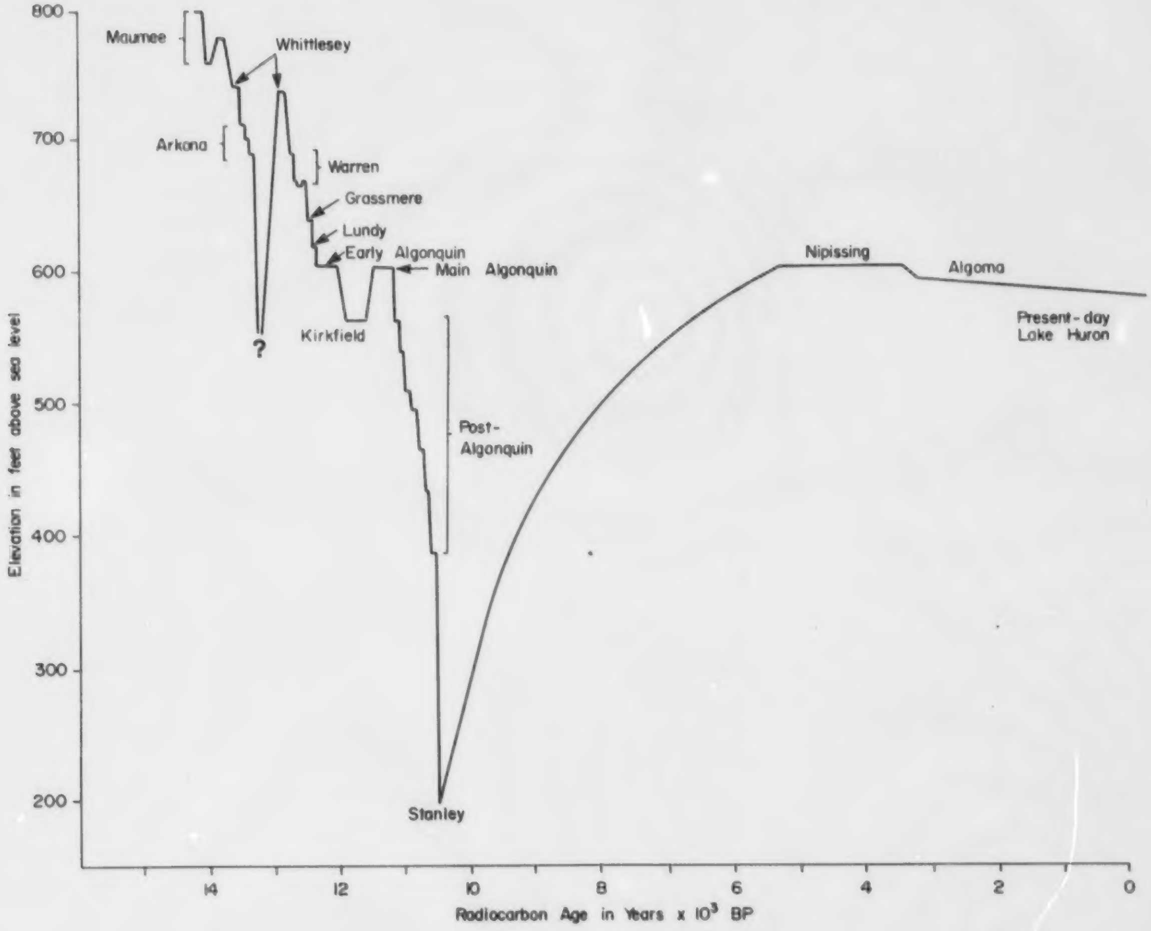


**Figure 21.43.** Ancestral lake phases in the Lake Huron basin (*modified from* Chapman and Putnam 1984).

Dummer moraine litter the surface. The matrix of the till is very sandy, usually containing lenses and discontinuous layers of sand. Ice-contact stratified drift deposits in kames and eskers occur in association with the melt-out and flowtills of this moraine.

The history of glacial Lake Algonquin that occupied parts of the basins of lakes Huron and Michigan is commonly divided into 4 parts (Eschman and Karrow 1985): 1) an Early Lake Algonquin southward-draining phase which occurred simultaneously with glacial Lake Schomberg and possibly after Schomberg waters had coalesced with those of Early Lake Algonquin; 2) a Kirkfield low water phase when the outlet changed to the east (Fenelon Falls outlet) allowing glacial meltwater to bypass the Lake Erie basin; 3) a Main Algonquin phase (high lake level) when drainage again became southward after the closing of the Kirkfield outlet; and 4) an Algonquin–Stanley phase of falling water levels which occurred as a series of lower outlets were uncovered to the northeast across Algonquin Park and into the Ottawa River valley.

Glacial Lake Iroquois occupied the western end of the Lake Ontario basin as Early Lake Algonquin existed in the lakes Huron and Michigan basins, about 12 300 years ago (Figure 21.44). Glacial Lake Iroquois drained southeastward via an outlet at Rome, New York. It expanded northeastward following the receding ice margin and flooded areas in the Trenton embayment and areas near Mazinaw Lake (Muller and Prest 1985). Several northward-diverging shorelines, produced as a result of isostatic recovery, are related to glacial Lake Iroquois in Ontario.

As the ice margin uncovered an outlet near Covey Hill, New York, the water level in the Lake Ontario basin fell to a level known as glacial Lake Frontenac. An outwash system and delta at Flinton, Ontario, graded to the Frontenac level, indicates that the glacier margin was not far from Mazinaw Lake during the Frontenac phase (Henderson 1973). Further ice retreat continued to open successively lower outlets. Proglacial lakes in the Champlain River valley coalesced with those in the Lake Ontario basin (glacial lakes Sydney, Belleville and Trenton).

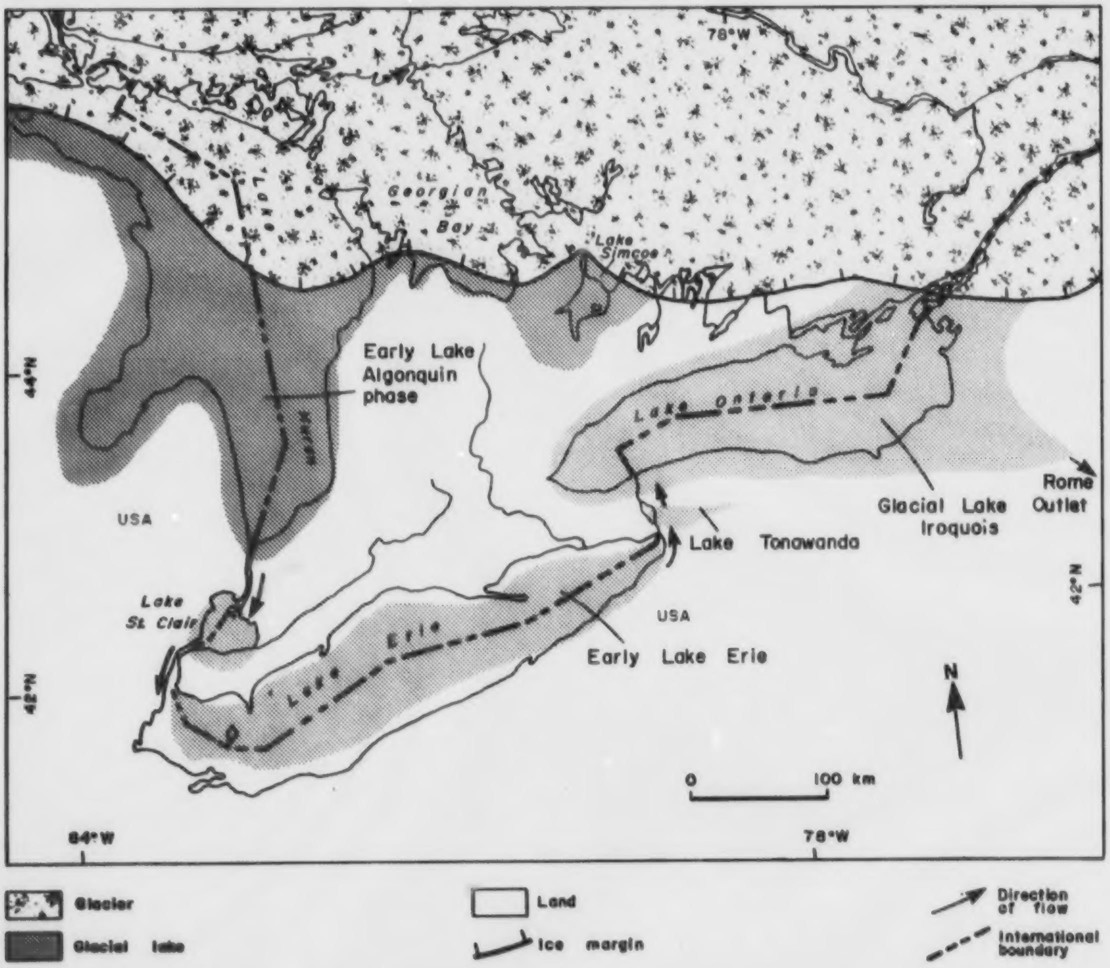

78°W
Lake Huron
Georgian Bay
Lake Simcoe
Early Lake Algonquin phase
44°N
USA
Lake Ontario
Rome Outlet
Glacial Lake Iroquois
Lake Tonawanda
42°N
Lake St. Clair
USA
Lake Erie
Early Lake Erie
42°N
N
0
100 km
84°W
78°W
Glacier
Glacial lake
Land
Ice margin
Direction of flow
International boundary

**Figure 21.44.** Glacial Lake Iroquois and Early Lake Algonquin (*modified from* Chapman and Putnam 1984).

The Kirkfield low water phase of glacial Lake Algonquin existed in the Lake Huron basin and drained along the Trent River system into glacial lakes Iroquois and post–Iroquois low levels, such as the Sydney, Belleville and Trenton levels (Gravenor 1957; Mirynech 1962). The Fenelon Falls outlet was in operation from about 12 000 to 11 300 years ago based on the relationship of the Kirkfield low water phase to the Two Creeks Forest Bed in the Wisconsinan (age about 11 900 years BP) and on the age of alluvial fill sequences along Lake Huron (Karrow et al. 1975; Miller et al. 1979). Isostatic uplift closed the outlet at Fenelon Falls, causing water levels in the lakes Huron and Michigan basins to begin to rise to the Main Algonquin level. Ponds and very low level lakes (Early Lake Erie) existed in the Lake Erie basin (Coakley and Lewis 1985).

The remarkable record of abandoned shoreline deposits and features, combined with the distribution of fine-grained glaciolacustrine sediments, provide evidence for the proglacial lake record in the Lake Ontario, Georgian Bay and Lake Huron basins.

## Later Events and Sediments

The Greatlakean Stade, which followed the Two Creeks Interstade approximately 11 800 years ago, was a period of glacial advance recorded by the Two Rivers Till in Wisconsin. In Wisconsin, the Two Creeks Forest Bed is overlain by glaciolacustrine sediments and the Two Rivers Till. This sequence of sediments indicates a rise in lake level preceding the overriding of the site by glacial ice. Although this advance is also recognized in the state of Michigan (Burgis and Eschman 1981), it has not yet been recognized in Ontario (Karrow 1984, 1989). It may be that the only significant ice advance occurred within the Lake Michigan drainage basin.

In Ontario, glacial deposits are thin and lacking in the southern part of the Canadian Shield where the evidence for such an advance would have been left. Very few recessional moraines have been recognized south of the Mattawa and French rivers. Mapping has failed to identify evidence for a significant readvance of the ice margin.

A readvance to the Lake Simcoe moraine is often suggested as a mechanism to close the Fenelon Falls outlet and to initiate the Main Algonquin phase (Karrow 1984), but field evidence in the outlet area does not support this suggestion (Finamore 1985). Water levels in the Lake Simcoe basin appear to have been approximately 15 to 20 m above the Main Algonquin level as the Lake Simcoe moraine was constructed, based on the evidence of a delta along the moraine at Logan Hill. Whether or not this high level relates to a level of glacial Lake Schomberg or Early Lake Algonquin has not been determined. Shorelines at a similar height above the Main Algonquin shoreline have been recognized on the west side of Lake Simcoe (Chapman and Putnam 1966) and on the Bruce Peninsula (Sharpe and Jamieson 1982), which may suggest the continuity of water levels in the lakes Huron and Simcoe basins prior to the opening of the Fenelon Falls outlet. The outlet must have closed as a result of isostatic rebound.

With continued northward recession of the Laurentide Ice Sheet margin in Ontario, glacial lakewaters in the Lake Ontario basin could expand into the Ottawa River valley almost as far north as Ottawa (Anderson et al. 1985; Anderson 1988; Rodrigues 1988). Glaciolacustrine rhythmites (varves) containing freshwater ostracodes occur beneath glaciomarine sediments of the Champlain Sea and overlie glacial and glaciofluvial deposits. With the uncovering of the St. Lawrence River valley approximately 11 700 to 11 500 years ago (Anderson 1988; Rodrigues 1988), water levels dropped in the Lake Ontario basin and allowed seawater to inundate the isostatically depressed Ottawa, Champlain and upper St. Lawrence river valleys (the western basin of the Champlain Sea).

An alternative and controversial mechanism by which the Champlain Sea formed and expanded up the Ottawa and St. Lawrence river valleys was proposed by Gadd (1980). Gadd suggested that a calving bay advancing up the Ottawa River valley allowed marine waters to enter the area around Ottawa first, while a remnant ice mass remained at the eastern end of the Lake Ontario basin, supporting higher level proglacial lakes (Lake Iroquois sequence of glacial lakes). This was suggested to help explain discrepancies between radiocarbon dates on marine shells from the Champlain Sea and younger dates on wood from glacial Lake Iroquois sediments in the Lake Ontario basin (Karrow 1981). The distribution of glaciofluvial landforms that occur in the Ottawa River valley and their internal structure and sedimentology, however, support general east-west oriented ice margins during deglaciation of the valley and not a calving bay (Gorrell, in prep.). Other authors have suggested that marine waters entered the Lake Ontario basin; however, this seems unlikely (Clark and Karrow 1984; Pair et al. 1985).

The general sequence of sediments deposited within the Champlain Sea includes: 1) ice-marginal delta and subaqueous fan sand and gravels (Rust and Romanelli 1975; Rust 1977, 1988; Sharpe 1988; Gorrell, in prep.); 2) laminated silt and clay containing trace fossils, present locally near unit 1 (Barnett 1987); 3) massive to colour-banded clayey silt to clay containing marine fossils; 4) laminated to varved silts and clays deposited as bottomsets of the large deltas formed during regression of the Champlain Sea; and 5) fossiliferous nearshore and beach sands and gravels.

The massive to colour-banded clays are by far the thickest (exceeding 50 m in places), most abundant and most widespread unit (Gadd 1977; Fransham and Gadd 1977). These fine-grained sediments are often referred to as "Leda" or "Quick" clays by engineers because of their behavioral characteristics when disturbed. Numerous landslides have occurred along slopes cut into these sediments. The laminated silts and clays associated with the large deltas that formed during the regression of the Champlain Sea are also very prone to landsliding. Large retrogressive slides have occurred within these laminated sediments; for example, the South Nation River landslide.

A large variety of fossils has been found in the marine sediments of the Champlain Sea including whales, seals and several species of invertebrates (Figure 21.45; *The Mercury,* November 30, 1977, p.A12; Harington 1988; Cronin 1988; Rodrigues 1988). Even fossil fish have been found within calcareous nodules near Ottawa (McAllister et al. 1988).

The closing of the Fenelon Falls outlet brought about the Main Algonquin phase of glacial Lake Algonquin. The outlet for the Main Algonquin phase was probably the Port Huron outlet at the south end of Lake Huron (Eschman and Karrow 1985); however, this has been questioned (Kaszycki 1985). The shoreline deposits and landforms of the Main Algonquin phase are very well developed around the lakes Michigan, Huron and Georgian Bay basins and the south-eastern end of the Superior basin (Cowan 1976, 1985; Futyma 1981). Probably this lake was relatively long-lived and the Fenelon Falls outlet was not used again (Karrow 1984).

As lower outlets began to open in the northeastern part of the lake basin across Algonquin Provincial Park (Chapman 1954; Harrison 1970; Ford and Geddes 1986), water levels of glacial Lake Algonquin began to fall in the lakes Michigan, Huron and the eastern end of the Superior basin; initiating the fourth phase of the history of glacial Lake Algonquin. The series of falling water levels, recorded by abandoned shoreline deposits and landforms, has been divided into 2 groups: an upper group of shorelines including the Ardtrea, Upper Orillia and Lower Orillia (Deane 1950); and a lower group including the Wyebridge, Penetang, Cedar Point, Payette, Sheguiandah and Korah (Stanley 1936). The ice margin had receded at least to Capreol, for an ice-marginal delta along the Cartier I moraine was built to the Payette level of glacial Lake Algonquin (Burwasser 1979; A. Heath, University of Waterloo, personal communications, 1990).

As Great Lakes drainage was being diverted through a series of channels across Algonquin Provincial Park, a lobe of glacial ice remained active in the Ottawa River valley. Water draining from glacial Lake Algonquin first entered the sea along the Bonnechere River, and as the ice margin uncovered lower outlets to the north, the water flowed via the Indian and Barron rivers and then the Petawawa River (Chapman 1954; Harrison 1970; Ford and Geddes 1986). The ice margin remained in contact with the Champlain Sea in Renfrew County during the early drainage events which used the Bonnechere and Indian rivers (Barnett 1988). The Petawawa delta formed along the Ottawa River, in an environment similar to a present-day Icelandic valley sandur, as the ice-lobe margin continued to recede. Downcutting and terracing of this delta began in response to lowering levels of the Champlain Sea, prior to the inflow of Great Lakes waters entering the sea via the Petawawa River (Barnett 1988).

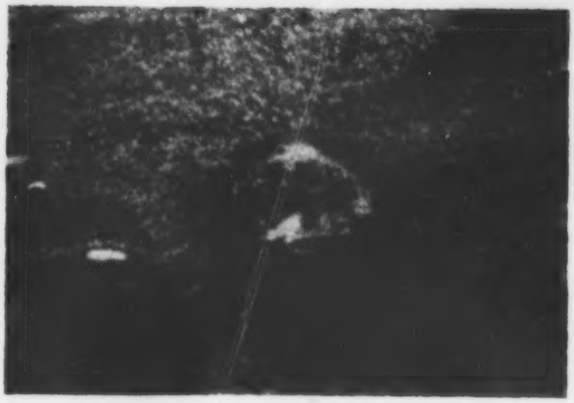


**Figure 21.45.** Barnacles from Champlain Sea sediments, near Ottawa. (Approximately 6 cm of the knife handle is showing.)

Glacial Lake Algonquin ended as the Mattawa and Ottawa river valleys became ice-free at about 10 400 to 10 000 years BP. Sediments deposited during the various phases of glacial Lake Algonquin contain a variety of fossil remains including pollen, diatoms, plant macrofossils, molluscs, insects and ostracodes (Eschman and Karrow 1985; Miller et al. 1985). The pollen assemblages indicate that glacial Lake Algonquin existed through the spruce-pine transition (approximately 10 400 years BP) or at least to shortly before 10 000 years ago (Eschman and Karrow 1985).

The Deux Rivières moraine, built across the Ottawa River valley by the receding ice margin, formed a drift dam to the water draining from the Georgian Bay basin (Figure 21.46a). The Aylen outlet across this dam controlled the water level of glacial Lake Barlow, which flooded areas along both the Mattawa River and the Ottawa River as far north as Lake Timiskaming (Vincent and Hardy 1979). Downcutting of the outlet at Aylen by about 75 m resulted in the draining of the part of glacial Lake Barlow in the Mattawa and Ottawa river valleys south of La Cave (Figure 21.46b). As a result, very low level lakes formed in the Georgian Bay (Lake Hough) and Lake Huron (Lake Stanley) basins. By this time, the shoreline of the Champlain Sea had regressed to the Ottawa area.

Glacial Lake Barlow was now controlled by an outlet at La Cave (Vincent and Hardy 1979). The outlet of glacial Lake Barlow quickly shifted to the rapids at Témiscaming, Quebec, and the Témiscaming phase of glacial Lake Barlow was initiated (Vincent and Hardy 1979). During the early part of this phase, approximately 10 000 years ago, the margin of the Laurentide Ice Sheet was at the Roulier moraine in Quebec and at its possible equivalent, the Chapleau II moraine (Sultan scarp) in Ontario. Local glacial lakes formed between the Chapleau II moraine and the drainage divide between Hudson Bay and the Great Lakes (glacial lakes Sultan, Oström and Ogilvie; Boissoneau 1967).

Within the basins of glacial lakes Barlow and Ojibway, the direction of ice flow changed from south-southwest to south-southeast during the deposition of the Matheson till (Veillette 1986). The Matheson till, a sandy silt to silty sand till, represents continuous ice cover over northeastern Ontario from at least the beginning of the Late Wisconsinan and possibly from the beginning of the Wisconsinan.

The Laurentide Ice Sheet still covered northwestern Ontario 11 300 years ago as the Kirkfield outlet was closing and the Main phase of glacial Lake Algonquin was beginning. The Lockhart phase of glacial Lake Agassiz was also beginning as the waters of glacial Lake Koochiching coalesced with the waters of the Cass phase of glacial Lake Agassiz in Minnesota, North Dakota and South Dakota (Figure 21.47). The margin of the Labrador Sector of the Laurentide Ice Sheet left the Rainy River basin shortly after. The Lake of the Woods–Rainy Lake moraine along the

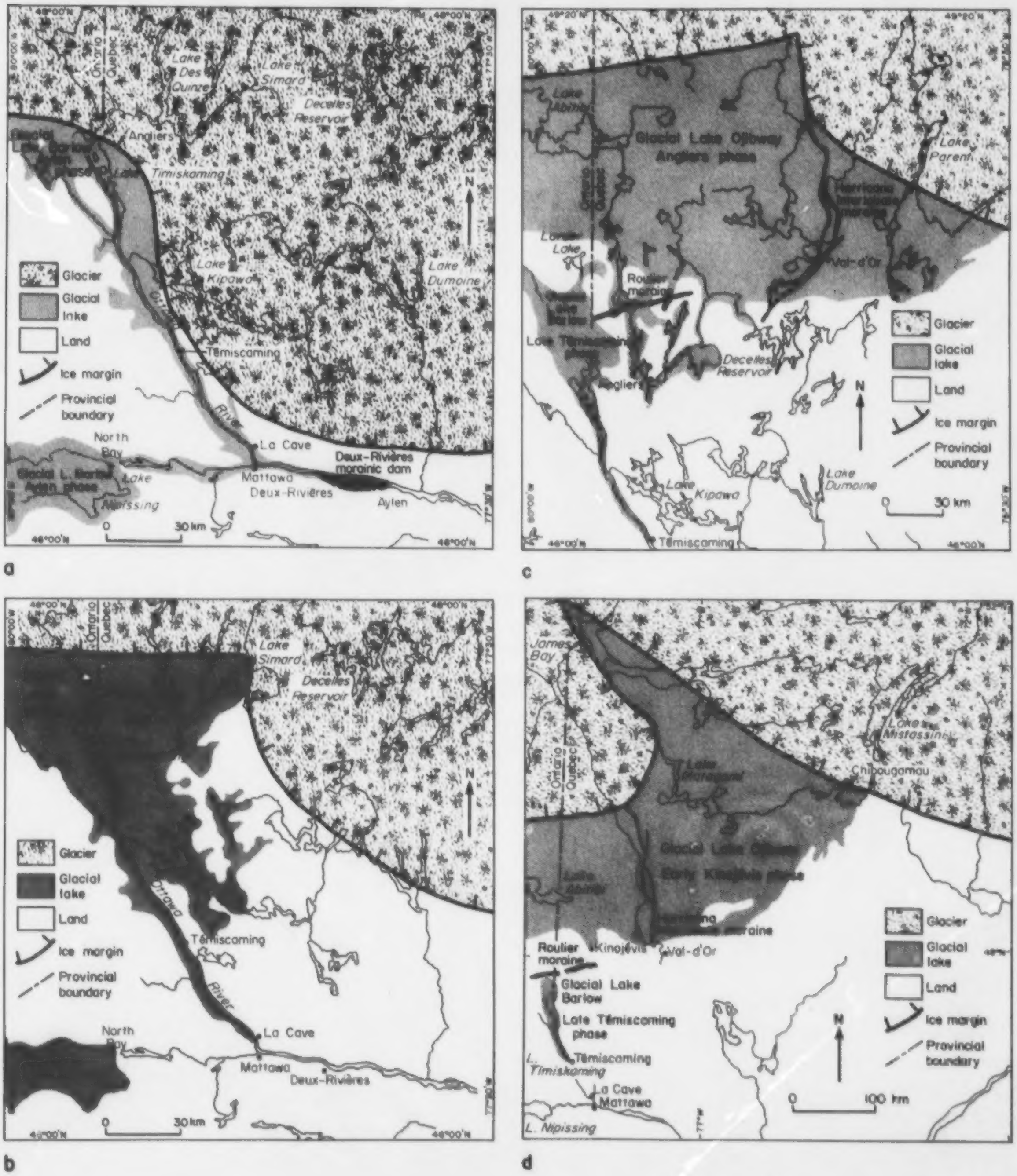

48°00'N
Ontario
Quebec
Lake Des Quinze
Lake Simard
Deceles Reservoir
Angliers
Glacial Lake Barlow Aylen phase
Lake Timiskaming
N
Lake Kipawa
Lake Dumoine
Glacier
Glacial lake
Land
Ice margin
Provincial boundary
Ottawa
Témiscaming
River
North Bay
La Cave
Deux-Rivières morainic dam
Glacial L. Barlow Aylen phase
Lake Nipissing
Mattawa
Deux-Rivières
Aylen
0
30 km
46°00'N
77°30'W
a
48°00'N
Ontario
Quebec
Lake Simard
Deceles Reservoir
N
Glacier
Glacial lake
Land
Ice margin
Provincial boundary
Ottawa
Témiscaming
River
North Bay
La Cave
Mattawa
Deux-Rivières
0
30 km
46°00'N
b
49°20'N
Lake Abitibi
Glacial Lake Ojibway Angliers phase
Ontario
Quebec
Lake Parent
Harricana Interlobate moraine
Lower Lake
Roulier moraine
Val-d'Or
Glacial Lake Barlow
Late Témiscaming phase
Decelles Reservoir
Angliers
Glacier
Glacial lake
Land
Ice margin
Provincial boundary
N
Lake Kipawa
Lake Dumoine
0
30 km
Témiscaming
46°00'N
c
James Bay
Lake Mistassini
Ontario
Quebec
Lake Matagami
Chibougamau
Glacial Lake Ojibway
Early Kinojévis phase
Lake Abitibi
Harricana moraine
Roulier moraine
Kinojévis
Val-d'Or
Glacial Lake Barlow
Late Témiscaming phase
Témiscaming
L. Timiskaming
La Cave
Mattawa
L. Nipissing
77°W
48°N
Glacier
Glacial lake
Land
Ice margin
Provincial boundary
N
0
100 km
d

**Figure 21.46.** Evolution of glacial lakes Barlow and Ojibway (*modified from* Vincent and Hardy 1979). a) Lake BarlowAylen phase; b) Lake Barlow-transition from the Aylen phase to the Témiscaming phase; c) Lake Ojibway-Angliers phase (Lake Barlow-late Témiscaming phase); and d) Lake Ojibway-early Kinojévis phase (Lake Barlow-late Témiscaming phase).

northern edge of the Rainy River basin formed during ice-marginal recession. This allowed waters of the Lockhart phase of glacial Lake Agassiz (or glacial Lake Johnston, Antevs 1951) to flood the basin and deposit glaciolacustrine sediments over the sandy till derived from the north-northeast. This till is essentially a noncalcareous, sandy silt to silty sand till. Clast content is variable ranging from 15 to 70% with little to no carbonate clasts. The till occurs as buried ground moraine in the Rainy River basin and is the surface till north of the basin. It is found associated with ice-contact stratified drift in the Lake of the Woods–Rainy Lake moraine (Figure 21.48).

Approximately 11 200 years ago, the margin of an ice lobe of the Keewatin Sector of the Laurentide Ice Sheet (the St. Louis sublobe) advanced up the Rainy River valley and deposited a clayey till (Whitemouth Lake Formation) and a sandy till (Marchand Formation; A.F. Bajc, Ontario Geological Survey, personal communication, 1991). The incorporation of glaciolacustrine sediments and carbonate rock fragments from the Paleozoic rocks of Manitoba produced the massive, calcareous, silt to silty clay Whitemouth Lake till, averaging 40 to 50% clay and 30 to 40% silt. Matrix carbonate content commonly exceeds 25% and clast content is usually less than 15%. About 50% or more of the clasts are carbonate rock fragments. The till is usually compact, blocky and vertically jointed. It is found mainly as ground moraine, but is commonly buried beneath glaciolacustrine sediments or organic-rich deposits.

The sandy Marchand till averages 35% sand and 40% silt. It is stonier and contains more matrix carbonate (averaging 35%) than the underlying Whitemouth Lake Formation (A.F. Bajc, Ontario Geological Survey, personal communication, 1991). The Marchand till also occurs mainly as ground moraine and is commonly buried beneath glaciolacustrine sediments or organic-rich deposits.

The Lockhart phase of glacial Lake Agassiz returned to the Rainy River area as the ice margins withdrew to the northeast (Labrador Sector) and to the west and northwest (Keewatin Sector). By about 11 000 years BP, the Eagle–Finlayson moraine was being constructed at the margin of the Laurentide Ice Sheet (*see* Figure 21.48). This sharp-crested, single ridge moraine that is composed of sand and gravel, rises over 30 m above the surrounding land and can be traced for over 240 km (Zoltai 1961). The southern end of the moraine was reported to have been overridden and to have been overlain by 2 m of till (Zoltai 1961). Subsequent investigations indicate that the till along the crest of the moraine is probably of flowtill origin and was deposited during moraine formation (Cowan 1987). The Steep Rock moraine and the Brule Creek moraine are possible southeastward extensions of the Eagle–Finlayson moraine.

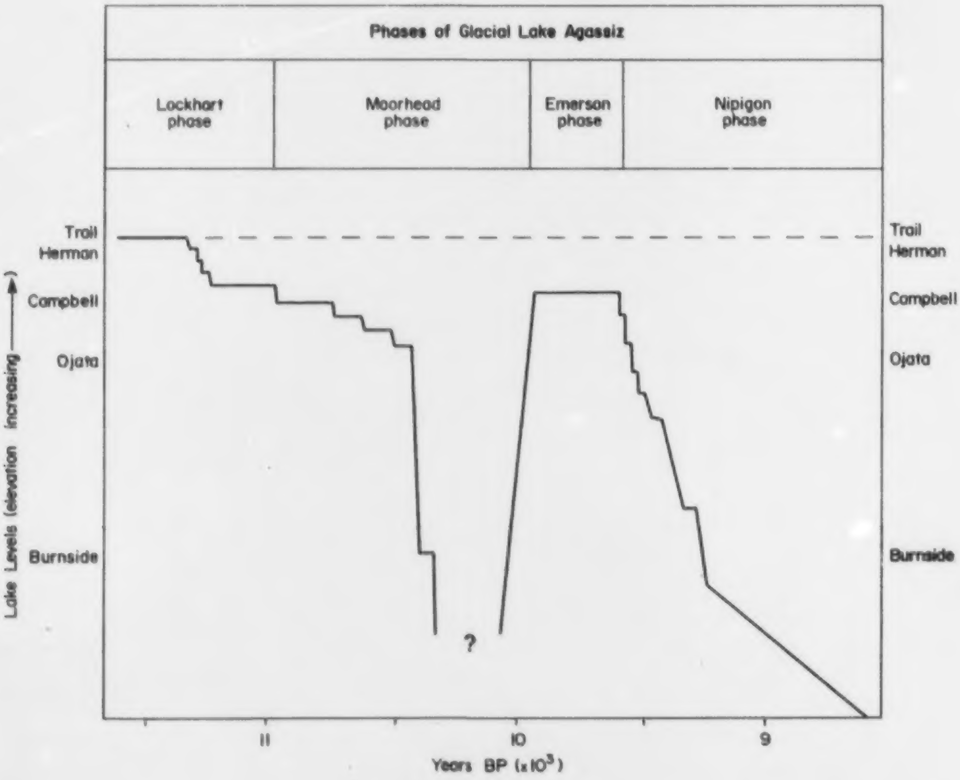


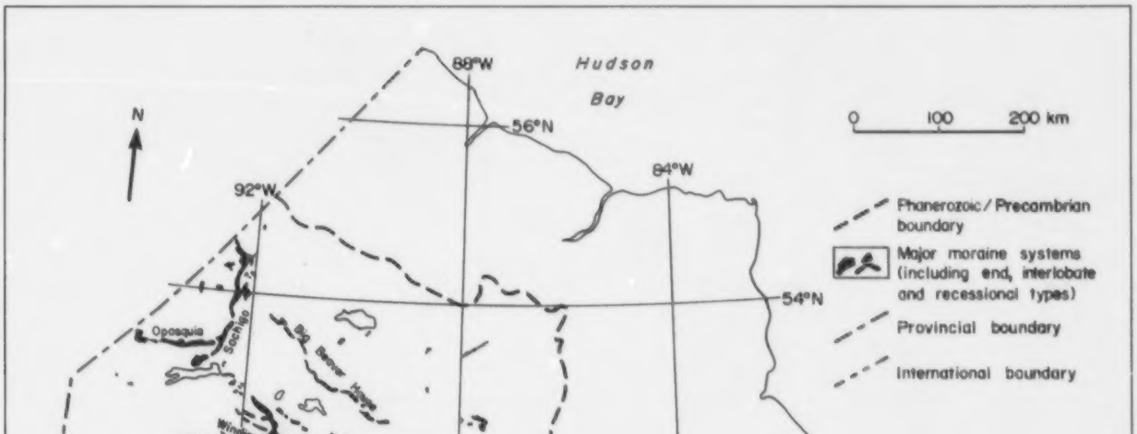
**Figure 21.47.** Phases of glacial Lake Agassiz (*modified from* Fenton et al. 1983).

Correlation with the Brule Creek moraine was favoured by Zoltai (1965).

The Hartmann and Lac Seul moraines roughly parallel the orientation of the Eagle–Finlayson moraine and are also prominent features of the landscape (*see* Figure 21.48). Composed of stratified sand and gravel of glaciofluvial origin, these moraines also rise over 30 m above the surrounding country. Although the moraines are thin, nearly continuous, long, linear ridges, their internal structure indicates deposition by currents flowing across the ridges from the northeast to southwest (Sharpe and Warman 1990). It has been suggested that these ridges formed as a result of a catastrophic release of meltwater (sheet flow) along the entire length of the ice margin, in response to a rapid lowering of the proglacial lakewaters as eastern outlets of glacial Lake Agassiz opened (Cowan and Sharpe 1988; Sharpe and Warman 1990; Sharpe and Cowan 1990). The rapid discharge may have been driven by the sudden increase in head differential between the piezometric surface in the glacier and the proglacial lakewater level.

Zoltai (1961), however, suggested that the Hartmann moraine formed during a readvance following a major retreat of the ice sheet margin. A readvance in the order of 30 km has been suggested for the younger Lac Seul moraine, based on the distribution of overridden varved clays (Zoltai 1965; Prest 1970). Prest (1970) correlated the Hartmann moraine with the Dog Lake and Marks moraines in the lakes Nipigon and Superior basins, but currently favours a correlation of the Marks moraine with the Nipigon moraine (Dyke and Prest 1987). Kaiashk interlobate moraine (Zoltai 1965) is an interlobate moraine built between the ice lobes that formed the Dog Lake and Hartmann moraines.

The Lockhart phase of glacial Lake Agassiz lasted until the ice margin retreated northeastward from the



88°W
Hudson
Bay
56°N
N
0 100 200 km
84°W
92°W
Phanerozoic / Precambrian boundary
Major moraine systems (including end, interlobate and recessional types)
54°N
Provincial boundary
International boundary
Opasquia
Sachigo
Big Beaver House
Windigo
Agutua
Miminiska
Manitoba
Ontario
Whitewater
Nakina
80°W
Crescent
Nakina
Oramon
50°N
Sioux Lookout
Hartmann
Lac Seul
Eagle–Finlayson
Kaiashk
Pinard
Quebec
Ontario
Lake of the Woods
Rainy L.
Nipigon
Dog Lake
Brule Creek
MacKenzie
Marks
Lake
Superior
USA
Chapleau
Chapleau
Chapleau
Cartier
Cartier
Cartier
Rawhide
Whiskey L.
McConnell L.
46°N

Figure 21.48. Moraines of northwestern Ontario (*modified from* Sado and Carswell 1987).

Eagle–Finlayson moraine approximately 10 800 years ago (Dredge and Cowan 1989). This recession opened lower outlets to the east into the Lake Superior basin along the Seine–Kashabowie and possibly Dog river valleys (Prest 1970; Teller and Thorleifson 1983). These valleys had been deglaciated by northward recession of the ice to Thunder Bay by this time (Dredge and Cowan 1989). The Moorhead low water phase of glacial Lake Agassiz then began and a large part of the southern part of the glacial Lake Agassiz basin, including the Rainy River area, became exposed. Ponds developed on this newly exposed landscape which quickly became vegetated.

Prest (1970) suggested that the readvance associated with the Dog Lake moraine closed off the eastern outlet along the Dog River, ending the first low water phase of glacial Lake Agassiz. Prest (1970) proposed 4 low water phases of glacial Lake Agassiz, where most workers suggest that there were only 2 low water phases (Clayton 1983; Teller and Thorleifson 1983; Teller 1987; Dredge and Cowan 1989). In Prest's (1970) version, following the readvance to the Dog Lake moraine that closed the Dog River outlet, the ice receded beyond the Sioux Lookout moraine, and the eastern outlets into Lake Nipigon opened, lowering the lake level. A readvance to the Lac Seul moraine, which was suggested by Zoltai (1965), closed these outlets and raised lake levels of glacial Lake Agassiz again (to the Campbell level). However, the evidence for such a readvance has been questioned (Cowan 1987). Ice-marginal retreat then reopened the eastern outlets and an advance to Sioux Lookout moraine closed them again, raising the level of glacial Lake Agassiz to the Campbell level for the last time.

The glacial history of this area, based on other workers' interpretations, calls for further ice-marginal retreat from the Eagle–Finlayson moraine to the vicinity of the Sioux Lookout moraine. A series of eastern outlets to Lake Superior via the Lake Nipigon drainage basin would then open, lowering the water level of glacial Lake Agassiz even further (Teller and Thorleifson 1983). The Moorhead low water phase ended about 10 000 years ago as the margin of the Laurentide Ice Sheet advanced to the Grand Marais moraines in northern Michigan and the Marks moraine in Ontario during the Marquette advance. Some workers correlate these moraines with the Dog Lake and Hartmann moraines in northwestern Ontario (Clayton 1983; Teller and Thorleifson 1983; Teller 1987; Dredge and Cowan 1989) and others to the Nipigon moraine (Dyke and Prest 1987). Glacial lakes of the Emerson (high water level) phase re-inundated areas that were drained during the Moorhead phase, burying the accumulations of organic material beneath glaciolacustrine sediments. The Marquette advance also deposited red clays in the Wabigoon and Seine river basins, producing up to 24 distinct red varves within the Emerson phase lake deposits.

Alternatively, Clayton (1983; Teller 1987) suggested that recession of the ice margin from the Eagle–Finlayson moraine did not open eastern outlets into the Lake Superior basin. Clayton proposed that the ice margin had to have receded much farther north before the eastern outlets into Lake Nipigon were uncovered. The Marquette advance in the Lake Superior basin and an ice advance to the Hartmann moraine initiated a series of higher levels in the glacial Lake Agassiz basin, referred to as the Emerson phase (Clayton 1983; Teller 1987).

Another explanation for the Moorhead low water phase is that lower outlets along the Clearwater Valley in Saskatchewan opened. Lake waters drained to the northwest and eventually down the Mackenzie River Valley in the Northwest Territories (Elson 1967; Christiansen 1979).

Most of these deglacial histories require extensive fluctuation of the ice margin across northwestern Ontario (Clayton 1983; Teller and Thorleifson 1983; Teller 1987). Some demand ice-marginal recession to the north to at least the Sioux Lookout or Agutua moraines and eastward to the Nipigon moraine to allow for eastward drainage into the Lake Nipigon drainage basin and then into the Lake Superior basin. This retreat is followed by a readvance to the Hartmann, Dog Lake and Marks moraines to deposit red clay into the glacial Lake Agassiz basin. Evidence for a readvance to the Hartmann moraine is lacking and over 800 km of ice front movement has to have occurred in less than 1000 years in these models.

Although rates of ice-marginal advance and retreat are enhanced when the glacier is situated within a lake basin, the glacial Lake Agassiz basin in Ontario is full of rock knobs and ridges that were islands in glacial Lake Agassiz (Zoltai 1961). These islands would have provided numerous pinning points (places where glaciers become grounded on and remain for some time) for the glacier and rates of ice-marginal retreat and advance would probably not be large. Ice streaming, by a deforming bed mechanism, in areas of thick till deposits in northern Ontario has also been suggested (Hicock 1988), though, the evidence presented to date is not conclusive. Whether or not localized ice streams would affect the response of the entire southern margin remains an unanswered question.

In addition, the deposits and landforms in this area, which include innumerable De Geer moraines, and well-integrated glaciofluvial systems in spatial association, argue against the rates of advance and retreat required in the above scenarios. An advance of an ice margin in the order of 200 to 300 km (to the Hartmann moraine) overriding glaciolacustrine sediments should produce fine-grained tills somewhere. These have not been reported in northwestern Ontario as of yet, except a till in the Lake Superior basin associated with the Marks moraine (Burwasser 1979). Studies of glaciolacustrine sediments in the Dryden area have failed to recognize any major or minor readvances of the ice margin (Sharpe and Warman 1990). The sediment sequences described represent ice-marginal recession or a deglaciation.

The crosscutting relationships of the end and interlobate moraines are also important. For example, the Nipigon moraine crosscuts and is, therefore, younger than the Kaiashk interlobate moraine, and the MacKenzie moraine appears to be younger than the Dog Lake moraine (*see* Figure 21.48). The MacKenzie moraine appears to correlate with the Marks moraine and possibly formed during the

Marquette advance. Another interesting landform relationship is the occurrence of De Geer moraines on the interfluves of the eastern outlet channels into the Lake Nipigon drainage basin (*see* Figure 11 *in* Telier and Thorleifson 1983). These drainage outlet channels must have been cut after the De Geer moraines were deposited, for De Geer moraines indicate that relatively deep proglacial lakes fronted the ice margin during deglaciation of this area. The ice-marginal position within the Lake Superior basin appears to have controlled the water level in the glacial Lake Agassiz basin and not the position of the northward receding (Rainy lobe) ice sheet margin as is suggested in many interpretations.

It is suggested here, that as the ice margin receded from the Eagle–Finlayson moraine, eastern outlets along the Seine–Kashabowie and possibly Dog river valleys opened, allowing glacial Lake Agassiz waters to drain into the Lake Superior basin, initiating the Moorhead low level phase of glacial Lake Agassiz (Teller and Thorleifson 1983; Dredge and Cowan 1989; Sharpe and Cowan 1990). The Eagle–Finlayson, Hartmann and Lac Seul moraines formed during deglaciation, possibly by the catastrophic release of meltwater in response to rapidly dropping water levels of glacial Lake Agassiz (Sharpe and Warman 1990). With farther northward recession, the Sioux Lookout and Whitewater moraines and the Kaiashk interlobate moraine formed. Water levels in front of the ice remained deep enough to form De Geer moraines along its receding grounding line. The area along the continental divide west of Lake Nipigon, where eastern outlet channels are located (Teller and Thorleifson 1983), remained below water level during deglaciation. These drainage channels became operative to lower water levels in the glacial Lake Agassiz basin after the ice margin in the Lake Superior basin had receded far enough north and east, to clear the Keweenaw Peninsula of Michigan. These channels may not have been in use, however, until after the Emerson phase, that is, only during the Nipigon phase following the Marquette advance (Dredge and Cowan 1989).

The Marquette advance to the Grand Marais moraines in northern Michigan, 10 000 years ago, raised water levels in the glacial Lake Agassiz basin to those of the Emerson phase (Clayton 1983; Drexler et al. 1983). In Ontario, this advance formed the Marks and MacKenzie moraines and possibly the Nipigon moraine. The Agutua moraine is the most likely correlative moraine and marks the end position of a small readvance of the Laurentide Ice Sheet margin ("Rainy River lobe") (Prest 1963). The major advance occurred within the Lake Superior and Lake Nipigon drainage basins. An ice advance filling the Lake Superior basin and extending to the Marks moraine could still allow the deposition of red clay into the Seine River and Wabigoon basins. Dispersal of the red clay would be confined by topography and, possibly, the sediment would be funnelled along pre-existing moraines. The above deglaciation scenario fails to explain the sequential south to north opening of the outlet channels west of Lake Nipigon, as has been suggested by Teller and Thorleifson (1983), and probably raises as many questions as it attempted to solve. Much more work is needed in this area to unravel the very complex history of deglaciation.

## HOLOCENE RECORD

The division between the Late Wisconsinan and the Holocene is somewhat arbitrary, in particular, with respect to the Ontario record. About half of the province remained covered by ice 10 000 years ago (Figure 21.49). It was the beginning of a major advance of the ice margin in the Lake Superior basin, the Marquette advance, not the end of a glaciation.

Low level lakes existed in the Lake Michigan (Lake Chippewa), Georgian Bay (Lake Hough), Lake Huron (Lake Stanley), Lake Erie (Early Lake Erie) and Lake Ontario (Early Lake Ontario) basins; the Champlain Sea had receded to east of Ottawa. The landscape of most of southern Ontario was covered by coniferous forests in which northern pines dominated over spruce, except in the more northern areas closer to the ice front (Webb et al. 1987). Deciduous trees such as oak were only beginning to become abundant, but only in the most southerly part of the province (Webb et al. 1987).

Man had entered the province for the first time. Early Paleo–Indian occupations between 11 500 to 10 200 years ago were associated with several of the abandoned lake shorelines. Fluted projectile points have been found along the shorelines of glacial lakes Whittlesey, Warren and Lundy in the Lake Erie basin (Deller 1976), along glacial Lake Algonquin shorelines (Storck 1984) in the Lake Huron basin and along shorelines of glacial Lake Iroquois in the Lake Ontario basin. Isolated surface finds of artifacts have been reported in the Champlain Sea basin (Mason 1960). Three fluted point complexes (Parkhill, Crowfield and Sydenham) have been identified from southern Ontario sites based on differences in site locations, tool kit contents, style of projectile points, types of chert used in tool manufacturing and differences in the methods used for producing flakes for tool manufacturing (Storck 1984).

The margin of the Laurentide Ice Sheet began its recession northward from the Grand Marais moraines, which mark the farthest extent of the Marquette advance in northern Michigan, shortly after the beginning of the Holocene. The ice margin in the Lake Superior basin ponded glacial Lake Duluth along its southwestern margin and glacial Lake Minong along its southeastern margin (Figure 21.50). The level of glacial Lake Minong was controlled by an outlet channel across Nadoway Point, Michigan (Farrand and Drexler 1985). With farther northward recession of the ice margin, a series of lower level lakes formed in the western end of the basin with outlets across northern Michigan.

Shortly before 9500 years ago, the ice margin receded north of the tip of the Keweenaw Peninsula allowing glacial Lake Minong to rapidly expand westward to the remaining deglaciated part of the Lake Superior basin. By 9500 years ago, glacial Lake Minong had expanded to its maximum size, flooding land along the north shore of Lake Superior. The lowering of glacial Lake Agassiz waters (Nipigon phase) and drainage eastward through the channels west of

Lake Nipigon had commenced, coincidentally, with ice-marginal retreat from the Keweenaw Peninsula and the Nipigon moraine. During deglaciation, the Onaman interlobate moraine, located northeast of Lake Nipigon, formed along a weakness within the glacier. Glacial Lake Kelvin formed along the ice margin in the Lake Nipigon drainage basin during the early stages of ice-marginal retreat (Figure 21.51).

Catastrophic floods flowing through the Lake Nipigon drainage basin and into the Lake Superior basin from glacial Lake Agassiz are thought to have cut down the drift sill at Nadoway Point in 5 or 6 steps, resulting in the post–Minong levels of the Lake Superior basin (Farrand and Drexler 1985). Once this outlet was cut down to the bedrock sill in the St. Marys River, water level in the Lake Superior basin stabilized at the low water Houghton level about 9000 years ago (*see* Figure 21.50).

A readvance of the Laurentide Ice Sheet margin to the Nakina moraine formed glacial Lake Nakina northeast of present-day Lake Nipigon, between the Great

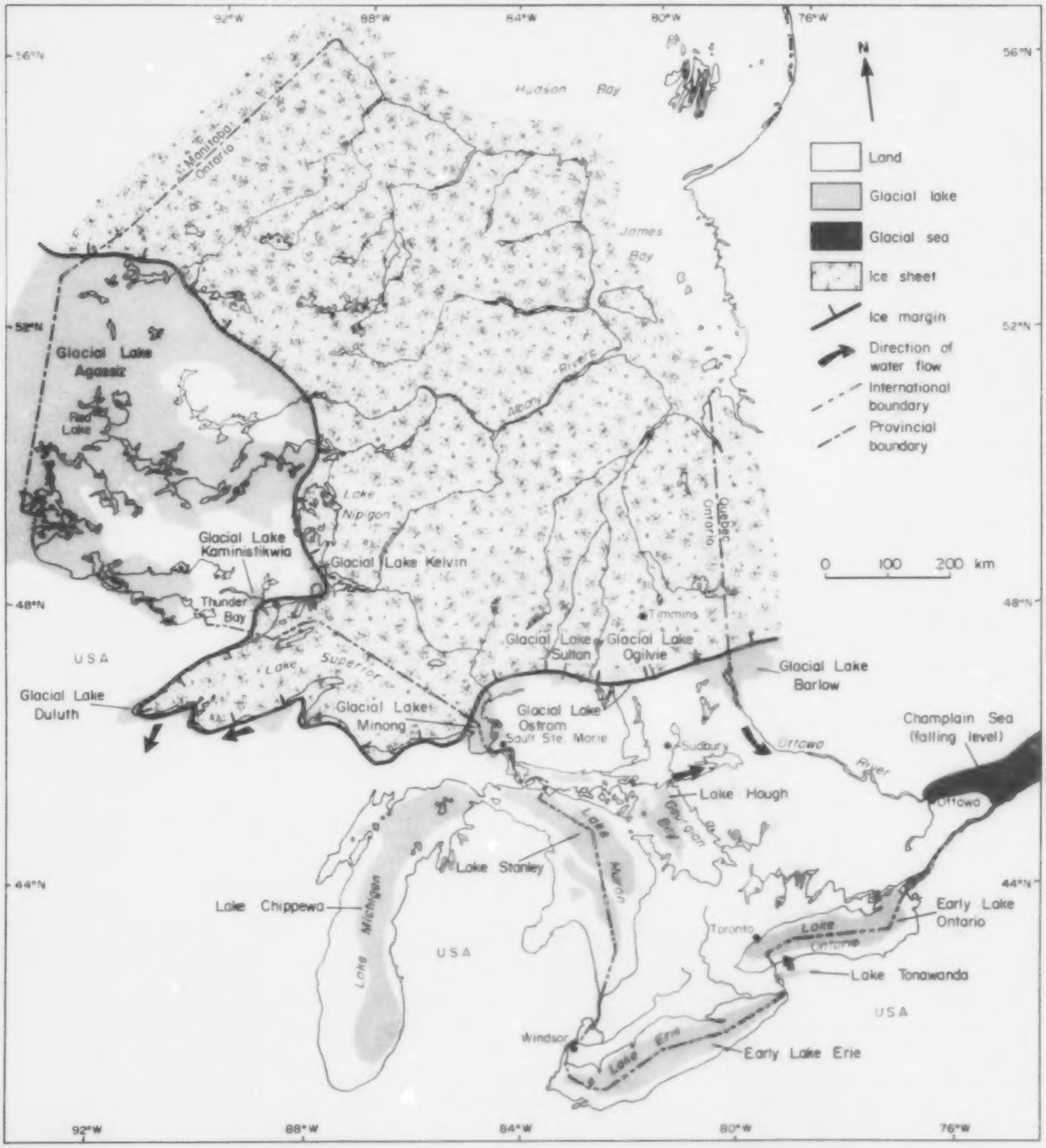

92°W
88°W
84°W
80°W
76°W
56°N
52°N
48°N
44°N
N
Hudson Bay
James Bay
Manitoba
Ontario
Land
Glacial lake
Glacial sea
Ice sheet
Ice margin
Direction of water flow
International boundary
Provincial boundary
Glacial Lake Agassiz
Red Lake
Albany River
Lake Nipigon
Quebec
Ontario
0 100 200 km
Glacial Lake Kaministikwia
Glacial Lake Kelvin
Thunder Bay
Timmins
USA
Lake Superior
Glacial Lake Sultan
Glacial Lake Ogilvie
Glacial Lake Barlow
Glacial Lake Duluth
Glacial Lake Minong
Glacial Lake Ostrom
Sault Ste. Marie
Sudbury
Ottawa River
Champlain Sea (falling level)
Ottawa
Lake Hough
Georgian Bay
Lake Huron
Lake Stanley
Lake Chippewa
Lake Michigan
USA
Toronto
Lake Ontario
Early Lake Ontario
Lake Tonawanda
USA
Windsor
Lake Erie
Early Lake Erie

**Figure 21.49.** Ontario at the beginning of the Holocene (various sources).

Lakes–Hudson Bay drainage divide and the Nakina moraine. Ice persisted long enough at the Nakina moraine to prevent glacial Lake Barlow–Ojibway and post–Minong lake levels from coalescing (Farrand and Drexler 1985).

Ice-marginal retreat from the Nakina and Agutua moraines probably permitted glacial lakes Agassiz and Ojibway to coalesce, thus ending the Nipigon phase of glacial Lake Agassiz and beginning the Ojibway phase (possibly correlative to the Gimli phase of Elson 1967; Dredge and Cowan 1989).

In northeastern Ontario, northward ice-marginal retreat from the Roulier moraine, north of Cobalt, allowed glacial Lake Barlow to expand northward and eastward and remain in contact with the ice margin. With the emergence of the Angliers sill as a result of isostatic rebound, Lake Barlow formed in the Temiskaming basin and glacial Lake Ojibway came into existence (*see* Figure 21.46c; Vincent and Hardy 1979). Further isostatic recovery shifted the controlling sill farther north to the Kinojévis sill and 2 Kinojévis phases of glacial Lake Ojibway formed (*see* Figure 21.46d). All 3 phases of Lake Ojibway were marked by rapid transgression of the lake northward on to newly deglaciated terrain while the outlets slowly shifted northward (Vincent and Hardy 1979).

As glacial lakes Ojibway and Agassiz coalesced, the entire southern margin of the Laurentide Ice Sheet in Ontario was in contact with standing water. Dredge and Cowan (1989) suggested that the ice margin began to calve and that former positions of grounding lines are marked by broad arcuate rises, curved in the opposite sense in regard to the curvature of the earlier deposited moraines. However, the Big Beaver House moraine, believed to have been developed during this phase of deglaciation (Dredge and Cowan 1989), is still arcuate in the same sense as the previously deposited Agutua moraine.

The Arnott moraine, west of Hearst, and the Pinard moraine have been interpreted as ice-contact deltas (Dredge and Cowan 1989). Prest (1970) suggested that the Nakina and Pinard moraines are correlative, but Dredge and Cowan (1989) suggested that the Pinard moraine may better correlate with the Big Beaver House moraine located in northwestern Ontario.

The margin of the Laurentide Ice Sheet in Ontario was becoming very unstable with such a large proglacial lake in contact with it. Extensive calving of icebergs is recorded by

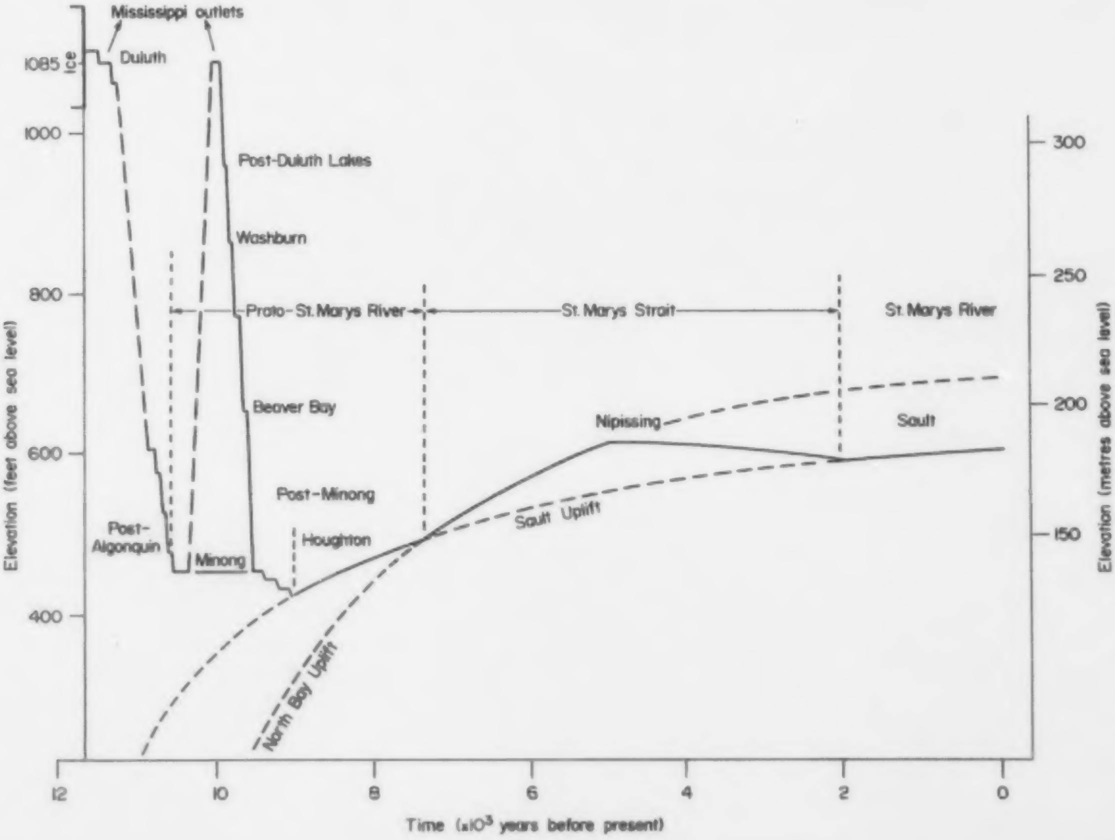


Figure 21.50. Ancestral lake phases in the Lake Superior basin (*modified from* Farrand and Drexler 1985). Curves represent the water level in the Lake Superior basin.

the numerous occurrences of iceberg scour marks within the glacial lakes Agassiz and Ojibway basins (Dredge and Cowan 1989). Irregular ice-marginal fluctuations have also been suggested, which have been interpreted as being surges that extended 50 to 75 km (Dredge and Cowan 1989). Based on varve sequences, individual fluctuations occurred within 25 years and are, therefore, considered surges (Hughes 1965; Dredge and Cowan 1989).

Fine-grained tills are associated with some of these surge events. The best documented events are the surges related to the Cochrane Till. Prest (1970) identified 2 phases of Cochrane advance, Cochrane I and II, which he thought extended some 800 to 1200 km. Dredge and Cowan (1989), however, have suggested that the Cochrane advances are a composite of many, much smaller surges along a continually retreating Laurentide Ice Sheet margin. The Cochrane advances are thought to have occurred between 8300 and 7900 years ago (Hughes 1965; Hardy 1982).

The Cochrane Till is a calcareous, clast-poor, silty clay to clay till. It averages approximately 40% clay, 45% silt and 15% sand in the area north of Timmins. Clast content is usually less than 3%. Matrix carbonate content ranges from 20 to 40%, with calcite being the dominant matrix carbonate mineral. It occurs as ground moraine on the surface or is buried beneath glaciolacustrine sediments of glacial Lake Ojibway, glaciomarine sediments of the Tyrrell Sea or organic accumulations of bogs and swamps.

A thick accumulation of rhythmically bedded silts and clays (varves) were deposited in glacial lakes Barlow, Ojibway and Agassiz (Figure 21.52). Some 2075 varves have been identified within the glacial Lake Barlow–Ojibway sediments (Hughes 1965). Hughes identified 3 thinning-upward cycles which he called the Lower, Frederick House and Connaught sequences. An abrupt change in thickness between the upper 2 cycles, the Frederick House and Connaught sequences, was related by Hughes (1965) to the Cochrane advance. Rhythmites deposited in glacial Lake Agassiz are widespread, but are concentrated within the low-lying areas between rock knobs and ridges in northwestern Ontario. In southern Manitoba, extensive clay plains occur as a result of this large proglacial lake. Clay plains in the Timmins and New Liskeard areas were formed during glacial lakes Barlow and Ojibway. Shoreline deposits of glacial lakes Agassiz, Barlow and

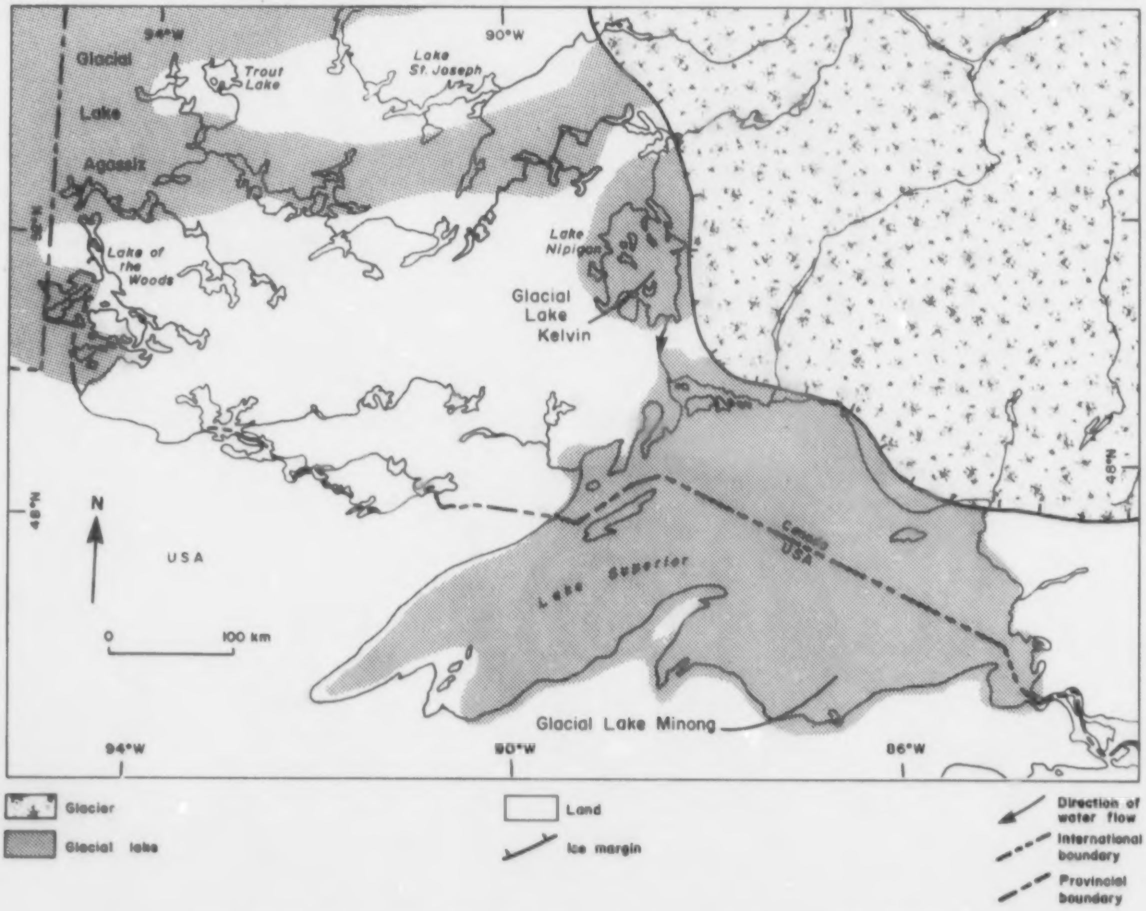

94°W
90°W
Glacial
Lake
Agassiz
Trout Lake
Lake St. Joseph
Lake of the Woods
Lake Nipigon
Glacial Lake Kelvin
49°N
48°N
48°N
N
USA
0
100 km
Lake Superior
Canada
USA
Glacial Lake Minong
94°W
90°W
86°W
Glacier
Glacial lake
Land
Ice margin
Direction of water flow
International boundary
Provincial boundary

**Figure 21.51.** Glacial lakes Agassiz, Kelvin and Minong in northwestern Ontario (*modified from* Clayton 1983).

**Figure 21.52.** Glacial Lake Barlow–Ojibway rhythmically bedded silts and clays (Frederick House varves), near Timmins (photograph *courtesy of* P.W. Alcock).

Ojibway are usually fragmented and best developed on positive relief landforms (eskers, kames and end and interlobate moraines) composed of icecontact stratified drift or till (Dredge and Cowan 1989).

About 8000 years ago, glacial Lake Ojibway and glacial Lake Agassiz catastrophically drained into Hudson Bay. Water level lowered drastically by some 300 m. The land surrounding Hudson Bay was immediately inundated, up to an elevation of about 145 m above mean sea level, by the Tyrrell Sea (Figure 21.53). A pebble gravel composed of pre-deposited glaciolacustrine rhythmite clasts marks this rapid drop and transgression of the Tyrrell Sea waters (Skinner 1973; Hardy 1977). The Laurentide Ice Sheet appears to have quickly disintegrated in Hudson Bay leaving active ice only in northern Quebec (Labrador Sector) and in Keewatin (Keewatin Sector) on the Canadian mainland.

Tyrrell Sea sediments consist predominantly of clays and silts that coarsen upward into nearshore and beach deposits of sand and gravel. Thicknesses of more than 7 m occur in the vicinities of the Pivabiska and Missinaibi rivers (Skinner 1973). Up to 60 m of glaciomarine sediments have been reported by Lee (1960) east of James Bay. Crustal rebound (isostatic recovery) produced a remarkable series of strandlines ringing Hudson Bay as the sea (Tyrrell Sea) regressed to its present location (Figure 21.54).

Peat blankets most of the Hudson Bay Lowland. Up to 4 m of peat have been observed, but, the cover is considerably thinner or absent near riverbanks and on raised shoreline deposits (Skinner 1973). Alluvium or fluvial sedimentation is confined to the bottom of steep canyon-like river valleys that have formed throughout the lowland. Skinner (1973) suggested the lack of lateral erosion in the area is probably the result of cohesive bank materials and isostatic rebound.

Isostatic rebound was affecting the water levels in the Great Lakes basins as well, during the Holocene. Following the opening of the North Bay outlet, water levels in lakes Michigan, Huron, Erie and Ontario were very low, with their shorelines almost entirely within the present lake shoreline. As the land surface became ice-free, it immediately began to rise isostatically, recovering from the weight of the ice sheet. As a result, water levels in the southern Great Lakes basins began to rise. Approximately 5000 years ago, as the outlet at North Bay rose differentially to the levels of the Chicago and Port Huron outlets at the south ends of lakes Michigan and Huron, respectively, the Nipissing Great Lakes came into existence in the basins of lakes Superior, Huron and Michigan and Georgian Bay (Figure 21.55). The Nipissing Great Lakes transgression produced strong shoreline features in Ontario that are some 30 m above the present shoreline of Lake Superior east of Thunder Bay (Farrand and Drexler 1985) and about 4 m above the southern Lake Huron shoreline. Erosion of the Port Huron outlet, which is cut into drift, continued while the erosion of the Chicago and North Bay outlets, both on rock, stopped, and these outlets were eventually abandoned. A series of falling levels, the most notable being Lake Algoma, formed as the water level fell to that of the present Upper Great Lakes. Erosion of the Port Huron outlet is often cited as the cause for these lower stages; however, climate-related fluctuations may explain some of the lower strandlines observed (Eschman and Karrow 1985). Water level in the Lake Erie basin also rose about 5 m above the present lake level some 4000 years ago as the result of a morainic dam at Fort Erie (the Fort Erie moraine) which had not yet been breached (Barnett 1985). The slow return of Upper Great Lakes drainage into the basin as a result of differential isostatic rebound, combined with the draining of Lake Tonawanda, a lake dammed between the Niagara and Onondaga escarpments, probably caused the cutting of the channel through the moraine down to rock in the Niagara River at Buffalo.

The Quaternary record in Ontario, although incomplete, is a record of major changes in climate and global conditions. It contains evidence of the terrestrial system's response to major shifts in climate and the interactions with the oceans and atmosphere. The records of warmer periods in the past provide analogies for future climatic warming and may, when combined with records from deep-sea (Martinson et al. 1987) or ice-sheet cores (Jouzel et al. 1989), form a more complete record against which numerical models used to predict global conditions in the future can be compared (Luckman 1989).

The effects of human activities on the global system have increased in magnitude significantly in the last 100 years to be detected on a global scale (Luckman 1989). Based on geological records, the world should be entering the next ice age; however, the build-up of carbon dioxide and other greenhouse gases produced by human activities is making the world unnaturally warm (Gribben 1989).

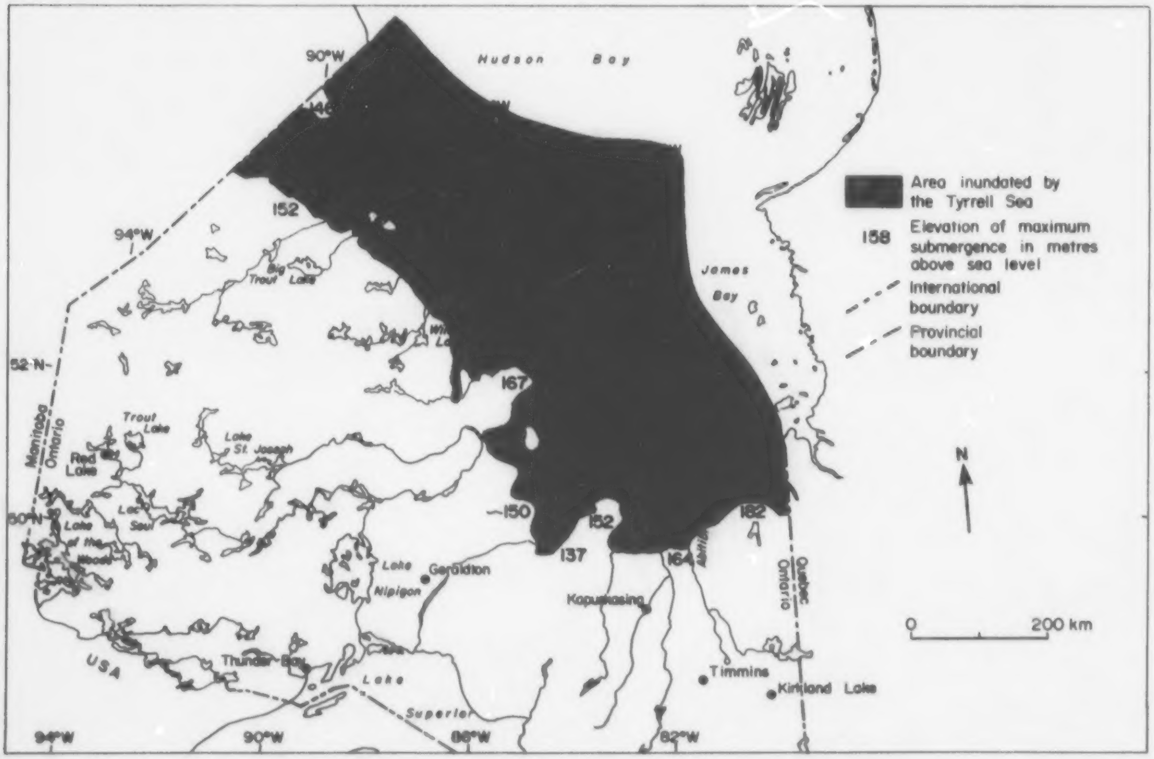
## NEOTECTONICS

Residual, high-horizontal stresses occur at relatively shallow depths in the bedrock in Ontario (White et al. 1973; Lo 1978). The stresses were first noted in excavations cut into rock, such as pipeline trenches and quarries, in which the walls of the excavation began closing or the quarry floors buckled (White et al. 1973; Lo 1978; Adams 1982). Natural features of stress release include: elongated compressional ridges or pop-ups (White et al. 1973; White and Russell 1982; Gorrell 1988); faulted, striated bedrock surfaces (Gorrell 1988); and displaced or disturbed shoreline features ("Day 8—Stop 39" by P.F. Finamore *in* Barnett and Kelly 1987). These natural features have formed since deglaciation (White and Russell 1982; Adams 1988). Many proposed origins of high-horizontal stresses in the bedrock have been proposed (Fakundiny et al. 1978). The 2 main theories are: 1) a response to glacial unloading (Adams 1988); and 2) regional tectonic forces (White and Russell 1982). A combination of both processes may be necessary to explain the occurrence of near-surface, high-horizontal stresses in the bedrock of Ontario.

## Summary of Events

The record of events that took place during the Quaternary Period in Ontario has been described in the previous sections of this chapter. Ontario was completely covered by the Laurentide Ice Sheet for the last time 20 000 years ago. The margin of the Laurentide Ice Sheet was extremely lobate in the Great Lakes region, responding to regional topographical conditions as well as the presence of water ponded in contact with the glacier margin. Proglacial ice-contact lakes formed and disappeared with fluctuations of the ice margin. The level of ponding was controlled by the lowest available outlet within the basin or adjacent basins. Glaciolacustrine sediments of gravel, sand, silt and clay were deposited within the series of proglacial lakes that followed the ice margin northward.

Advances of ice over glacial lake (glaciolacustrine) sediments generally produced fine-grained tills around the basins of the Great Lakes. Fine-grained till also occurs in the Rainy River area (glacial Lake Agassiz basin) and in the area around Hudson Bay where glaciolacustrine sediments were also overridden.



90°W
Hudson Bay
148
152
94°W
Big Trout Lake
James Bay
Area inundated by the Tyrrell Sea
158 Elevation of maximum submergence in metres above sea level
International boundary
Provincial boundary
52°N
167
Trout Lake
Lake St. Joseph
Red Lake
Manitoba
Ontario
N
50°N
Lake of the Woods
Lac Seul
150
152
182
A
137
164
Abitibi
Lake Nipigon
Geraldton
Kapuskasing
Quebec
Ontario
0
200 km
USA
Thunder Bay
Lake Superior
Timmins
Kirkland Lake
94°W
90°W
86°W
82°W

Figure 21.53. Distribution of the Tyrrell Sea (*modified from* Dredge and Cowan 1989).

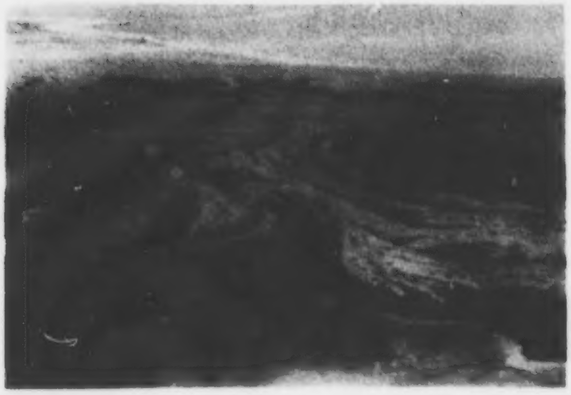


**Figure 21.54.** Abandonned Tyrrell Sea shorelines, Cape Lookout, Polar Bear Provincial Park (photograph *courtesy of* R.G. Gorman).

Several stages in the deglaciation of Ontario are depicted in a sequence of illustrations (Figure 21.56). Many attempts to reconstruct the configuration of the Laurentide Ice Sheet margin during deglaciation have been made, particularly in relation to the evolution of the Great Lakes (for example, Hough 1958 and Prest 1970, and in the papers *in* Teller and Clayton 1983; Fulton 1984a, 1989a; Karrow and Calkin 1985; Fulton and Andrews 1987; Gadd 1988). These previous reconstructions should be consulted for alternative views.

# APPLIED ASPECTS OF QUATERNARY DEPOSITS

## Introduction

The Laurentide Ice Sheet left a variety of deposits and landforms throughout Ontario. This legacy has been both a blessing and a hindrance during the development of the province of Ontario. The thick cover of Quaternary deposits in the southern part of the province, combined with a favourable climate, has produced soils which support a wide range of crop types. The Laurentide Ice Sheet has provided materials used in the construction of roads and buildings, and has also provided natural storage areas for water. Only a few of the applied aspects of the Quaternary geology of Ontario are mentioned below.

## Agricultural and Forestry Soils

The soils in Ontario are for the most part developed in glacially eroded and transported debris, and are not residual soils. They are comparatively very young. Soil-forming processes could only begin once the landscape had become uncovered by the Laurentide Ice Sheet or after the draining of the many proglacial lakes or seas that inundated large parts of Ontario. The oldest soil is probably less than 18 000 years old, while most of the province's soils are considerably younger. The soils that have formed are relatively thin and are well supplied with nutrients.

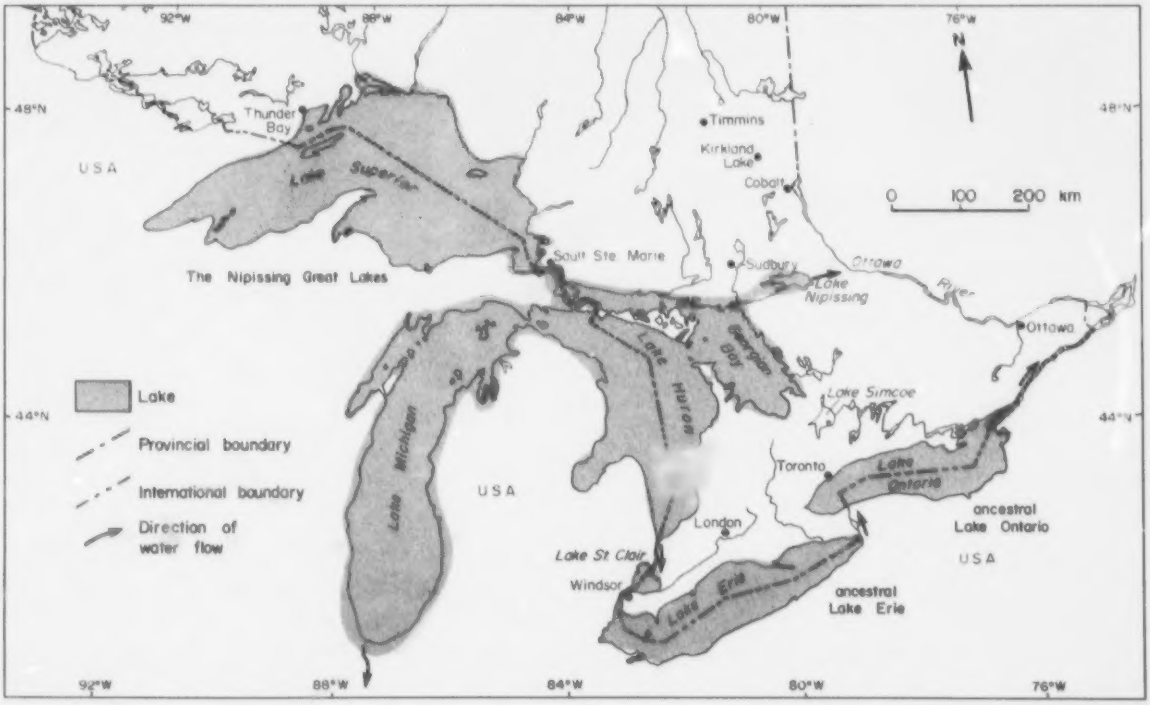
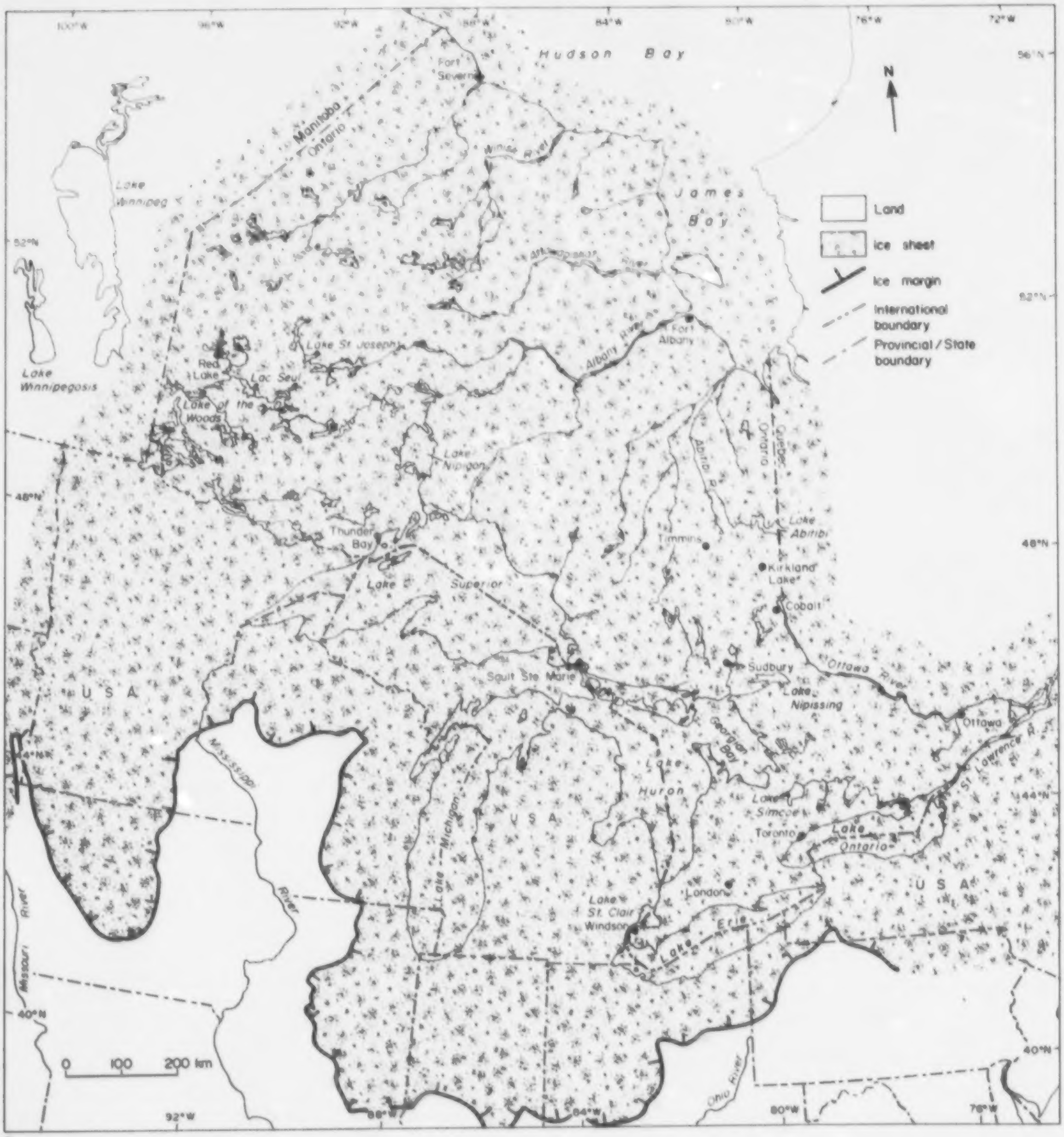

92°W
88°W
84°W
80°W
76°W
48°N
44°N
N
Thunder Bay
USA
Lake Superior
Timmins
Kirkland Lake
Cobalt
0
100
200 km
Sault Ste. Marie
Sudbury
Lake Nipissing
Ottawa River
Ottawa
The Nipissing Great Lakes
Lake Huron
Georgian Bay
Lake Simcoe
Lake
Provincial boundary
International boundary
Direction of water flow
Lake Michigan
USA
Toronto
Lake Ontario
ancestral Lake Ontario
London
Lake St. Clair
Windsor
Lake Erie
ancestral Lake Erie
USA

**Figure 21.55.** The Nipissing Great Lakes (*modified from* Prest 1970).

Many of the properties of the soils in Ontario are inherited from the rocks and sediments or organic accumulation that they are developed on (parent material of the soil). The parent material's particle size distribution, mineralogy and age combined with climactic, hydrologic and topographic factors help determine the soil type that will develop (Acton 1989a).

The amount of clay in the parent material determines the water retention capability, which influences the type of native vegetation or the agricultural capability (Acton 1989a). Soils developed on coarser-textured parent material, like sand and gravel, have lower water retention capabilities; in dry areas, the agricultural capability will be low. Soils developed on finer-grained parent materials, such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Lake Winnipeg
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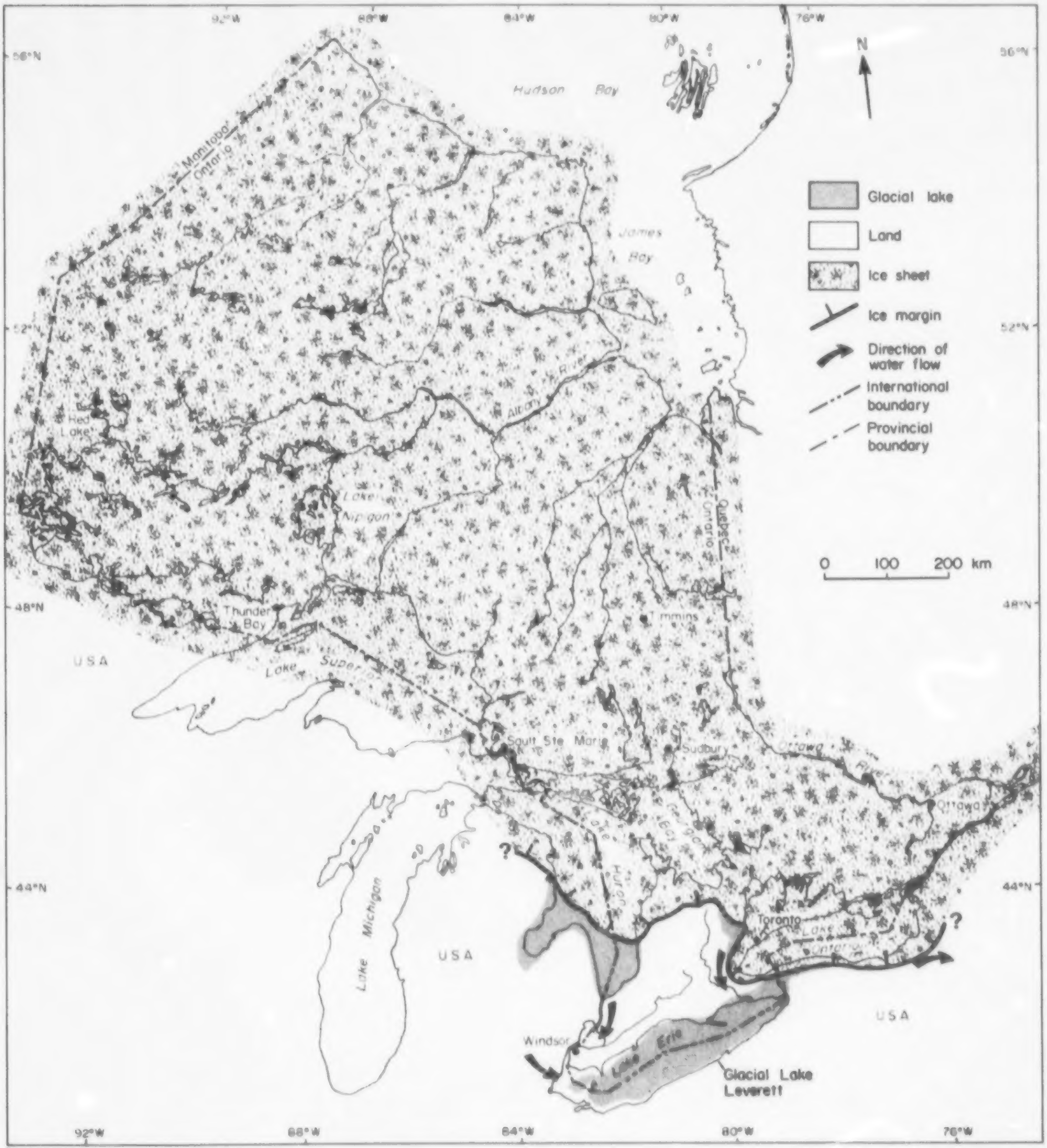
Attawapiskat River
Albany River
Fort Albany
Lake St Joseph
Red Lake
Lac Seul
Lake Winnipegosis
Lake of the Woods
Lake Nipigon
Abitibi R
Ontario
Quebec
Lake Abitibi
Timmins
Kirkland Lake
Cobalt
Thunder Bay
Lake Superior
Sault Ste Marie
Sudbury
Ottawa River
Lake Nipissing
Ottawa
St Lawrence R.
U S A
Mississippi River
Georgian Bay
Lake Huron
Lake Michigan
U S A
Lake Simcoe
Toronto
Lake Ontario
U S A
London
Lake St. Clair
Windsor
Lake Erie
Missouri River
Ohio River
0
100
200 km
N
Land
ice sheet
ice margin
International boundary
Provincial / State boundary
a

**Figure 21.56a.** Deglaciation of Ontario: the Nissouri Stade glacier maximum about 18 000 years ago. Many sources were used in the construction of the paleogeographic map presented, except where noted.

as glaciolacustrine clays and silts, will retain greater amounts of water and in dry areas will be the preferred soil types for agriculture. In wetter areas, however, the fine-grained soils will retain water, and because of their low permeability will be poorly aerated, possibly flooded and may be less desirable for agriculture (Acton 1989a).

Mineralogy of the soil affects the ion exchange capacity and nutrient levels; thus, the productivity of the soils. In general, sandy soils have low ion exchange capacities and fine-grained soils have a larger exchange capacity and a greater capacity to supply nutrients. The mineralogy of the soil, however, affects the amount of nutrients available for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Lake Nipigon
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Ontario
Sault Ste Marie
Sudbury
Ottawa River
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Georgian Bay
Lake Huron
Lake Michigan
USA
Toronto
Lake Ontario
USA
Windsor
Lake Erie
Glacial Lake Leverett
?
?
Glacial lake
Land
Ice sheet
Ice margin
Direction of water flow
International boundary
Provincial boundary
0 100 200 km

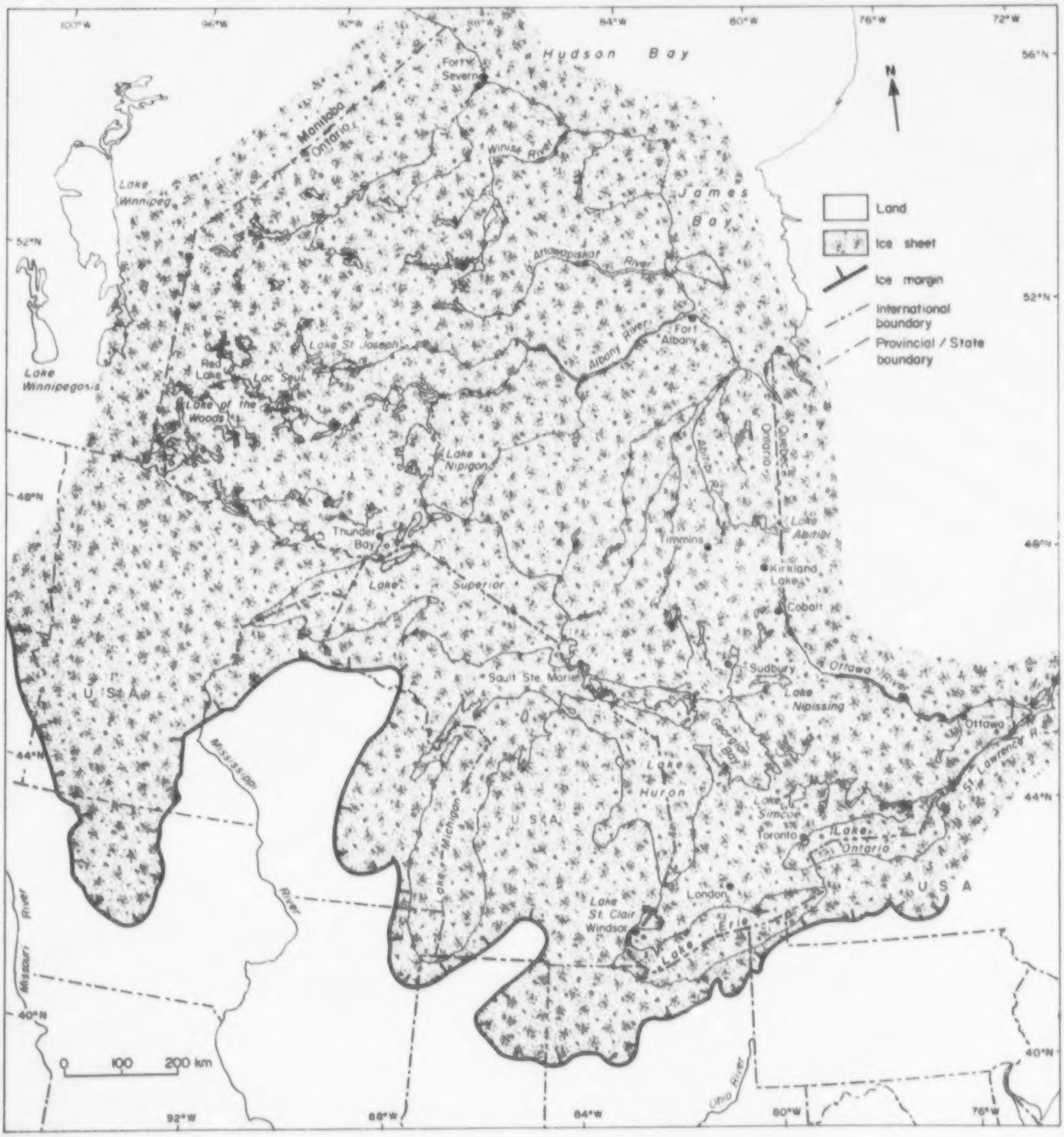
b

**Figure 21.56b.** Deglaciation of Ontario: maximum retreat during the Erie Interstade approximately 15 500 years ago. Many sources were used in the construction of the paleogeographic map presented, except where noted.

exchange. Sandy soils developed on till derived from granites will probably not have nutrient levels as high as sandy soils developed on tills derived from mafic igneous or metavolcanic rocks (van Groenewoud and Ruitenberg 1982). Productivity will likely be greater in the nutrient-rich soils, which may be an important consideration in reforestation project planning.

The southern part of Ontario has fertile soils which combined with favourable climatic conditions enable a large variety of field, vegetable and fruit crops to be grown (Presant 1989). Extensive vegetable production occurs in some of the better drained organic soils, such as in the Holland Marsh. Luvisolic soils are the dominant soil type. Gleysolic soils have developed in the poorly drained areas underlain by fine-grained glaciolacustrine sediments and brunosolic soils have developed in the stony areas of shallow till near the Canadian Shield (Presant 1989; *see* Webber 1975 for characteristics of the various soil types).

In the northern parts of Ontario, podzolic soils are abundant on well and imperfectly drained areas. Agriculture is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Manitoba
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Winisk River
Lake Winnipeg
James Bay
Attawapiskat River
Albany River
Fort Albany
Lake St Joseph
Red Lake
Lac Seul
Lake Winnipegosis
Lake of the Woods
Lake Nipigon
Ontario
Quebec
Abitibi R.
Lake Abitibi
Timmins
Kirkland Lake
Cobalt
Thunder Bay
Lake Superior
Sault Ste. Marie
Sudbury
Ottawa River
Lake Nipissing
Ottawa
St. Lawrence R.
Georgian Bay
Lake Huron
Lake Simcoe
Toronto
Lake Ontario
U.S.A.
Mississippi River
Lake Michigan
London
Lake St. Clair
Windsor
Lake Erie
Missouri River
Ohio River
0 100 200 km
N
Land
Ice sheet
Ice margin
International boundary
Provincial / State boundary

c

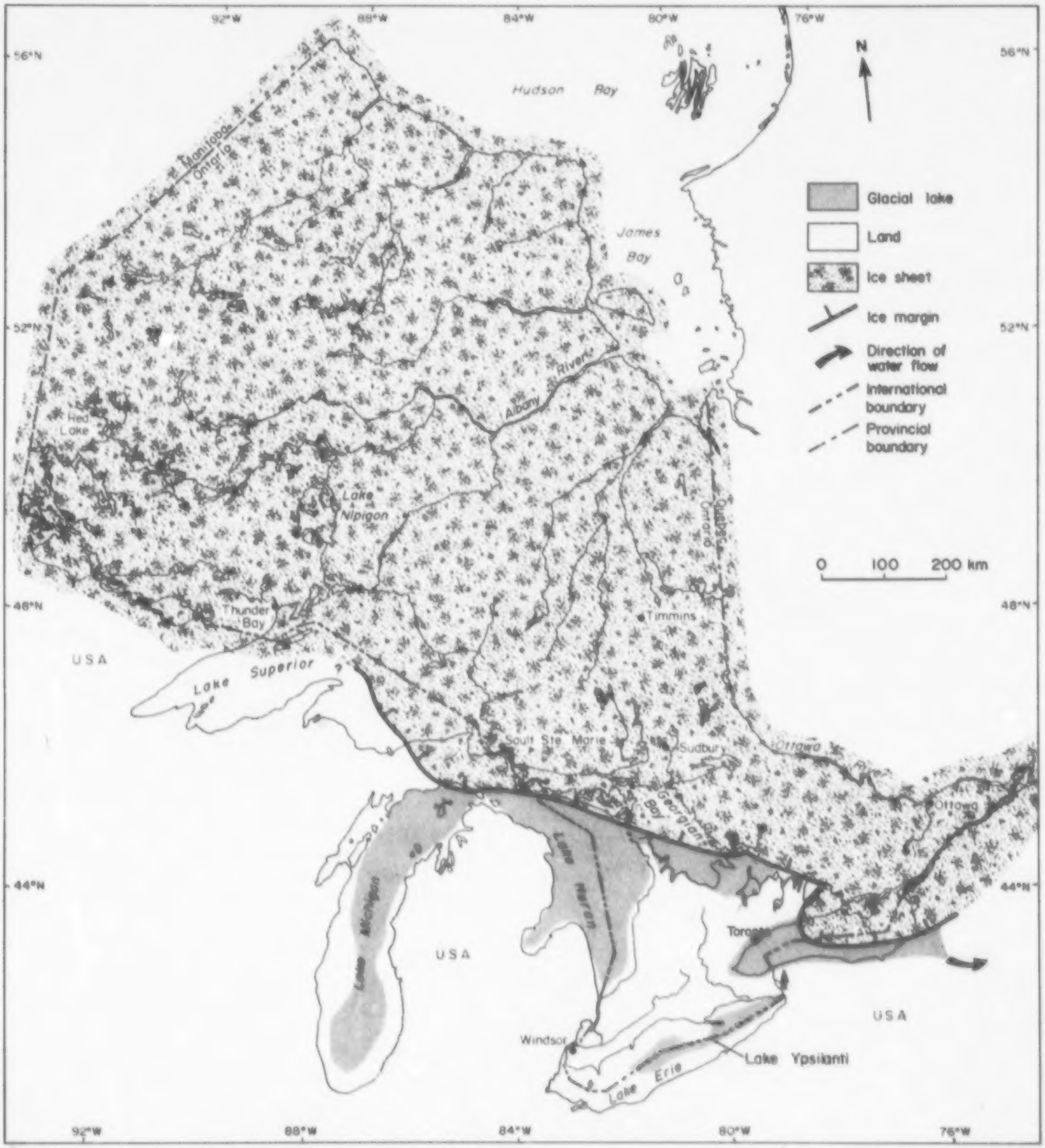
**Figure 21.56c.** Deglaciation of Ontario: the Port Bruce Stade glacier maximum about 14 800 years ago. Many sources were used in the construction of the paleogeographic map presented, except where noted.

essentially restricted to the large areas of glaciolacustrine deposits in the former glacial lake basins such as glacial lakes Agassiz, Barlow and Ojibway. Agriculture is also practised along the southern margin of the Canadian Shield on glaciofluvial, glaciolacustrine and glaciomarine deposits (Acton 1989b).

The relatively young soils of Ontario are thin and sensitive to erosion if disturbed and not properly managed. Topsoil in areas of sandy soils is particularly susceptible to wind erosion (Figure 21.57).

## Engineering Geology

Industrial growth and urban development have required large quantities of construction materials and the need to understand the distribution, engineering characteristics and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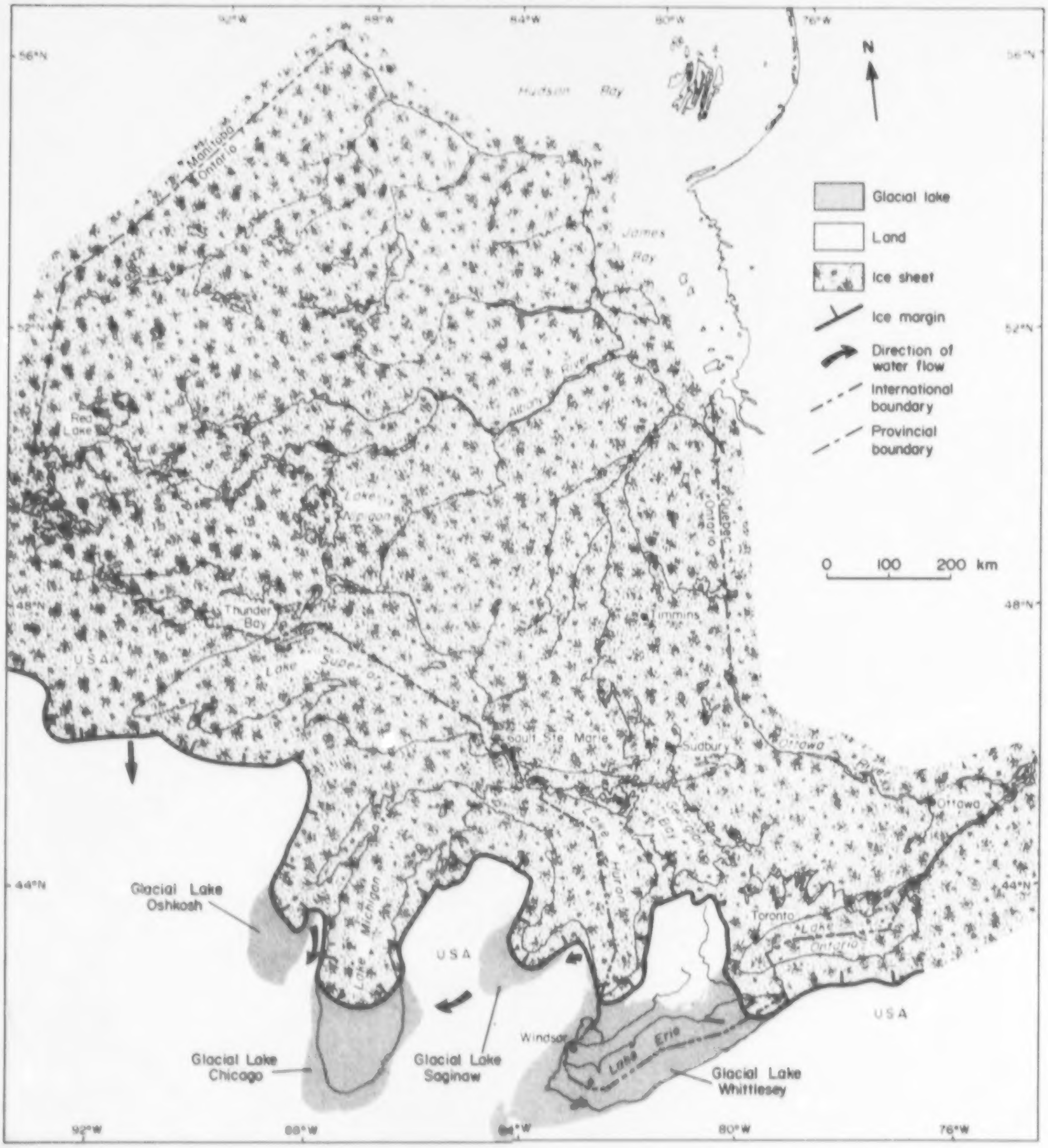
Ontario
Red Lake
Albany River
Lake Nipigon
Quebec
Ontario
Thunder Bay
Timmins
USA
Lake Superior
Sault Ste. Marie
Sudbury
Ottawa R.
Ottawa
Georgian Bay
Lake Huron
Lake Michigan
Toronto
USA
USA
Windsor
Lake Erie
Lake Ypsilanti
Glacial lake
Land
Ice sheet
Ice margin
Direction of water flow
International boundary
Provincial boundary
0 100 200 km
d

**Figure 21.56d.** Deglaciation of Ontario: maximum retreat conditions during the Mackinaw Interstade about 13 200 years ago. Many sources were used in the construction of the paleogeographic map presented, except where noted.

behavior of the Quaternary materials and bedrock beneath our feet.

## FOUNDATION AND EXCAVATIONS

The characteristics of the materials encountered during excavation will determine a variety of factors including: the stability of the walls of the excavation; the type of shoring required; if groundwater drainage or flooding will be a problem, the most suitable excavation methodology to use; and the cost and timing of excavation (Mathewson 1981). For the cut-and-fill projects, often required for highways, runways and rail beds, information on the characteristics of the material to be encountered is also needed. This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N
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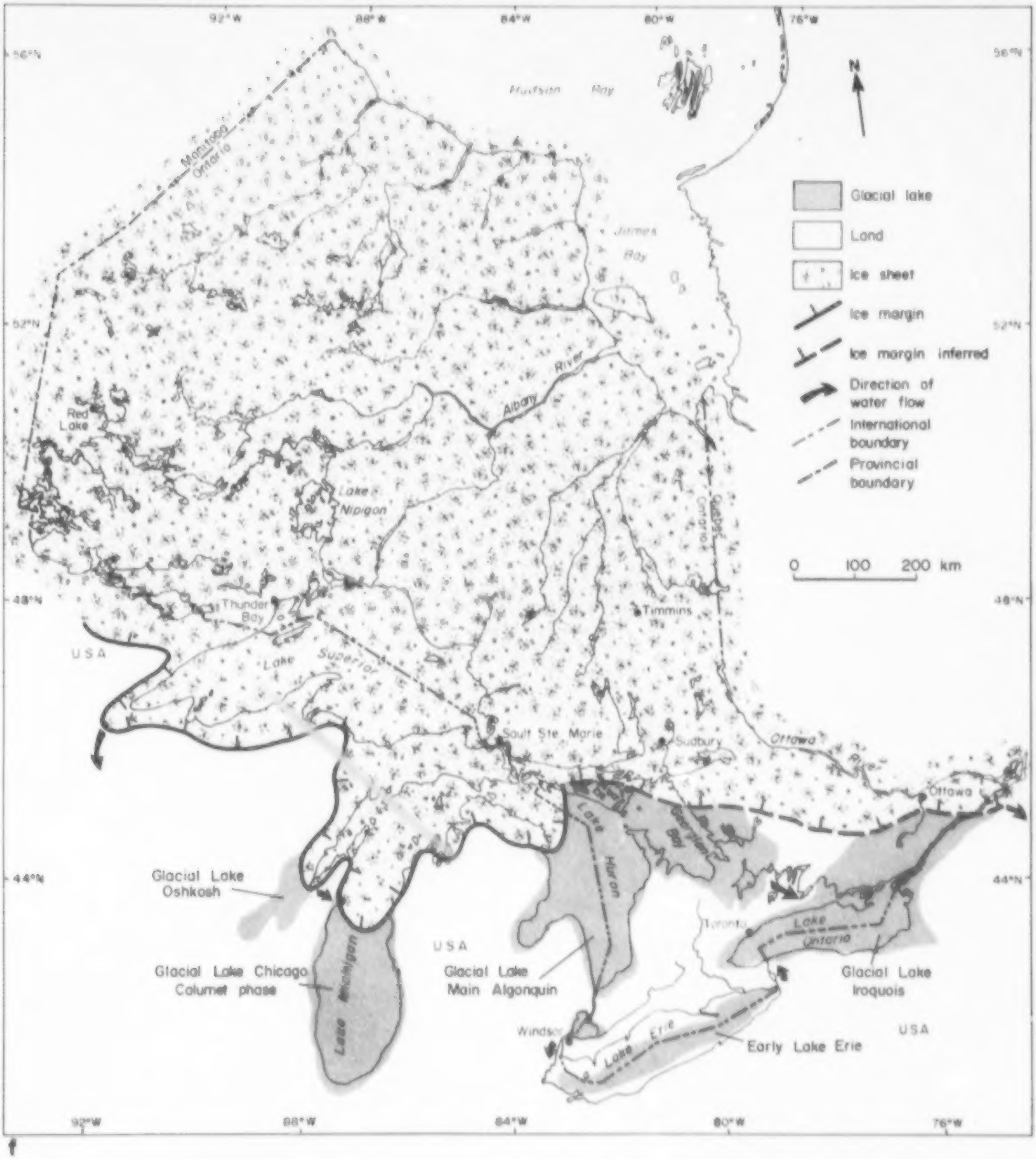
Manitoba
Ontario
James Bay
Albany River
Red Lake
Lake Nipigon
Thunder Bay
USA
Lake Superior
Timmins
Ontario
Quebec
Sault Ste. Marie
Sudbury
Ottawa River
Ottawa
Lake Michigan
Lake Huron
Georgian Bay
Toronto
Lake Ontario
USA
USA
Windsor
Lake Erie
Glacial Lake Oshkosh
Glacial Lake Chicago
Glacial Lake Saginaw
Glacial Lake Whittlesey
Glacial lake
Land
Ice sheet
Ice margin
Direction of water flow
International boundary
Provincial boundary
0
100
200 km

6

**Figure 21.56e.** Deglaciation of Ontario: the Port Huron Stade glacier maximum 13 000 years ago. Many sources were used in the construction of the paleogeographic map presented, except where noted.

information is used to decide whether a natural stable slope can be produced or whether slope retention or support structures are necessary. The materials removed from the cut are often used for fill and have to be evaluated for this use. The cost of transporting material for fill usually precludes selective borrowing and long distance transport of fill material (Mathewson 1981). The bearing capacities of the Quaternary materials and bedrock are considered when planning foundations. Site geology is a controlling factor in the location and types of large-scale construction projects that can be undertaken.

In areas of thin drift cover, i.e., the bedrock-dominated areas, construction costs can be much higher because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Hudson Bay
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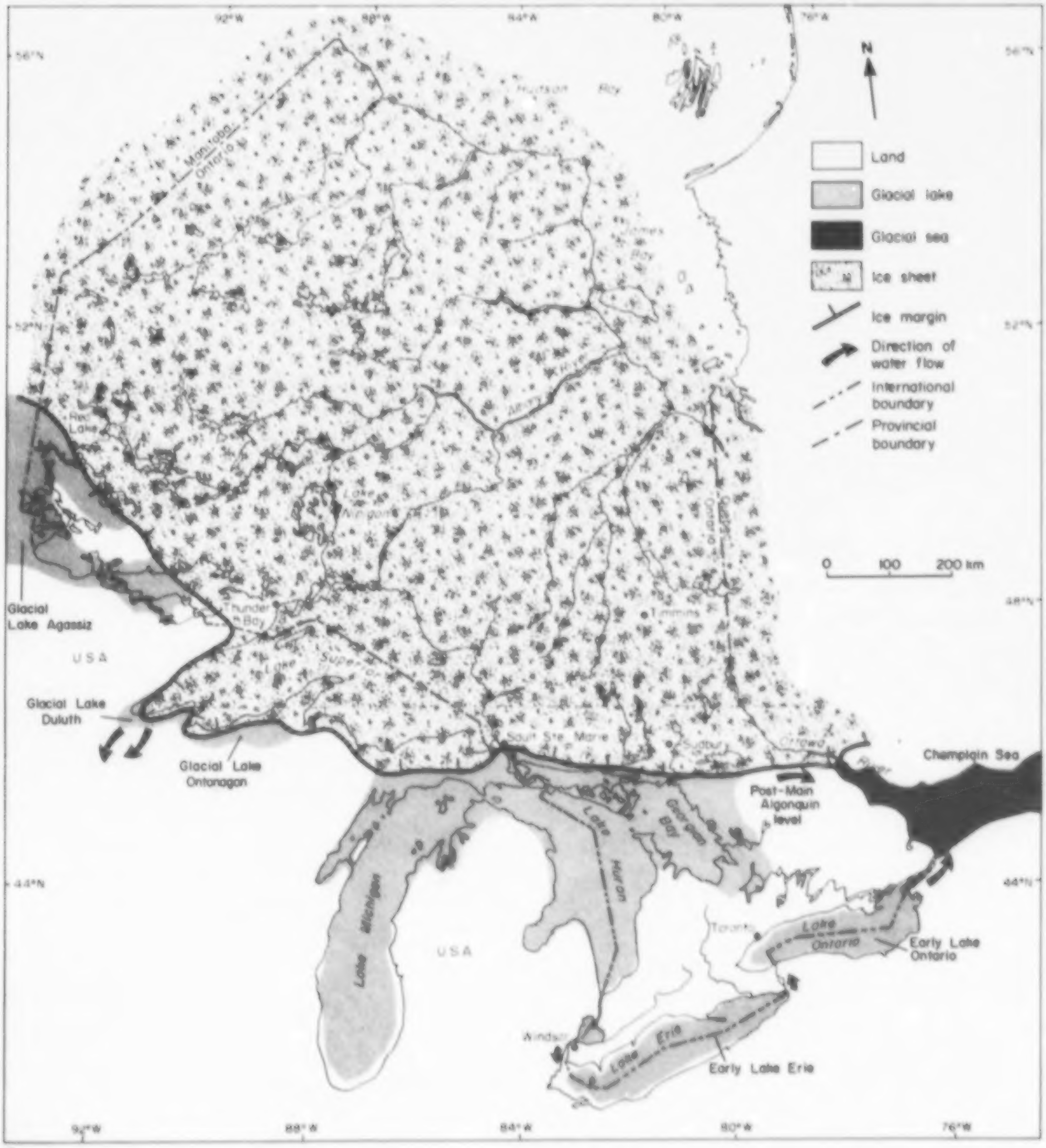
Ontario
James Bay
Glacial lake
Land
Ice sheet
Ice margin
Ice margin inferred
Direction of water flow
International boundary
Provincial boundary
Red Lake
River
Albany
Lake Nipigon
Quebec
Ontario
0 100 200 km
Thunder Bay
Timmins
USA
Lake Superior
Sault Ste. Marie
Sudbury
Ottawa River
Ottawa
Lake Huron
Georgian Bay
Glacial Lake Oshkosh
Lake Michigan
USA
Toronto
Lake Ontario
Glacial Lake Chicago Calumet phase
Glacial Lake Main Algonquin
Glacial Lake Iroquois
Windsor
Lake Erie
Early Lake Erie
USA
92°W
88°W
84°W
80°W
76°W
f

**Figure 21.56f.** Deglaciation of Ontario: the Greatlakean Stade glacier maximum about 11 800 years ago. Many sources were used in the construction of the paleogeographic map presented, except where noted.

drilling and blasting of rock is often required and tills are usually bouldery and stony. The coarse-textured till and sand and gravel deposits on the Canadian Shield are generally suited for most types of construction and often provide well-drained foundation sites (Gartner et al. 1981). Irregular or steep slopes associated with ice-contact and pitted outwash deposits may pose some problems for construction. High water tables and compressible soils such as peat must also be considered. Areas of bogs, swamps and modern alluvium (fluvial deposits on flood plains of present-day creeks and rivers) should be avoided for general construction because of periodic flooding, high water tables and compressive soils (Gartner et al. 1981).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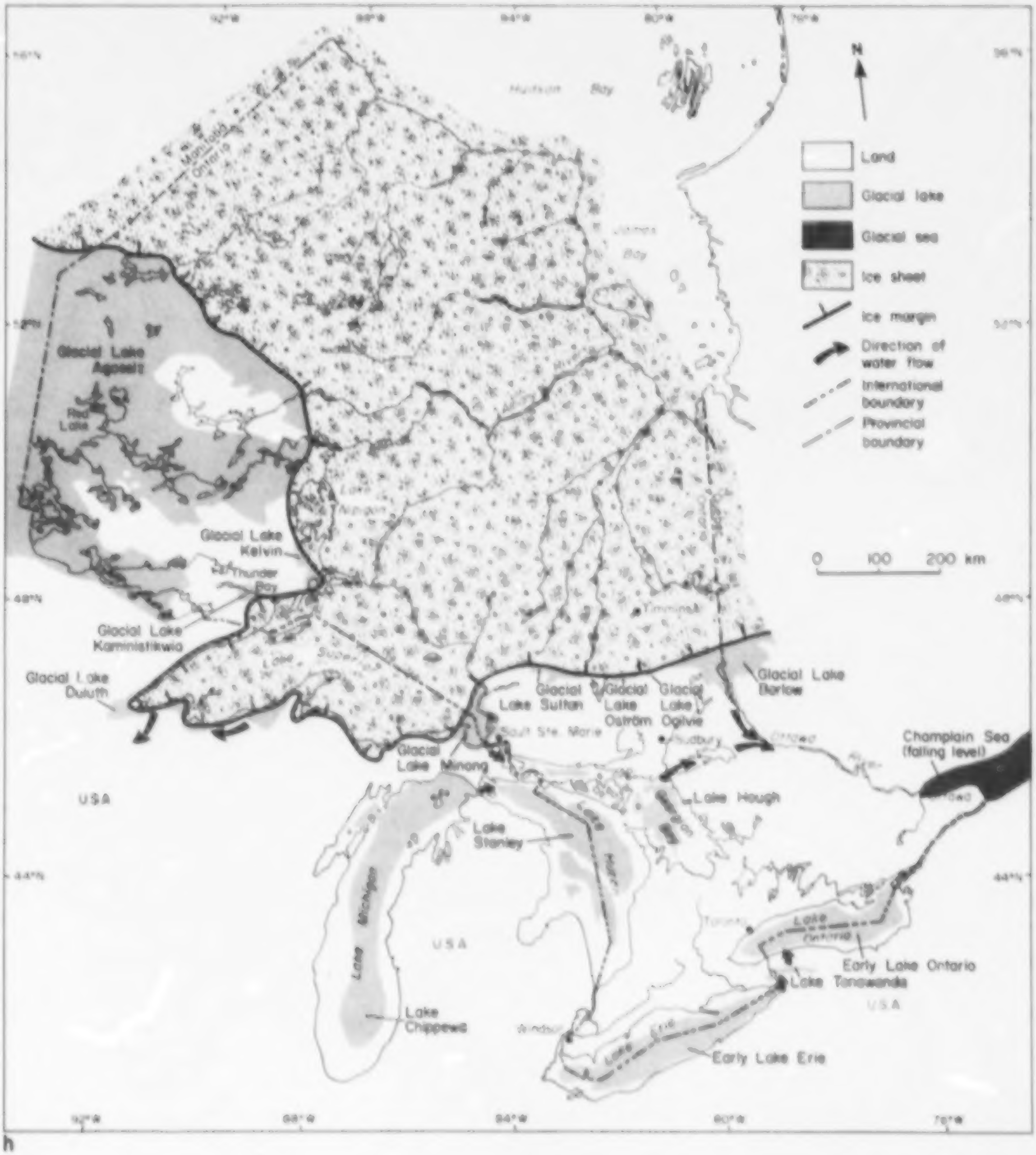
James Bay
Manitoba
Ontario
Quebec
Ontario
Red Lake
Lake Nipigon
Albany River
Thunder Bay
Timmins
Lake Superior
Sault Ste. Marie
Sudbury
Ottawa River
Land
Glacial lake
Glacial sea
Ice sheet
Ice margin
Direction of water flow
International boundary
Provincial boundary
0
100
200 km
Glacial Lake Agassiz
USA
Glacial Lake Duluth
Glacial Lake Ontonagon
Champlain Sea
Post-Main Algonquin level
Lake Huron
Georgian Bay
Lake Michigan
USA
Toronto
Lake Ontario
Early Lake Ontario
Windsor
Lake Erie
Early Lake Erie

g

**Figure 21.56g.** Deglaciation of Ontario: approximately 11 000 years ago during the maximum extent of the Champlain Sea. Many sources were used in the construction of the paleogeographic map presented, except where noted.

In areas of Ontario where the cover of Quaternary sediments is thick, construction is relatively easy and inexpensive (Karrow 1989). Problems encountered include: the requirements for cut and fill associated with the topography of some glacial landforms; drainage and high water tables; saturated sands and silts; compressive soils; and difficulties in excavating in tills that are stony, bouldery or highly overconsolidated (Karrow 1989).

Ice-contact stratified drift deposits are typically good reservoirs for groundwater. The coarse texture and positive relief of these deposits enable them to act as groundwater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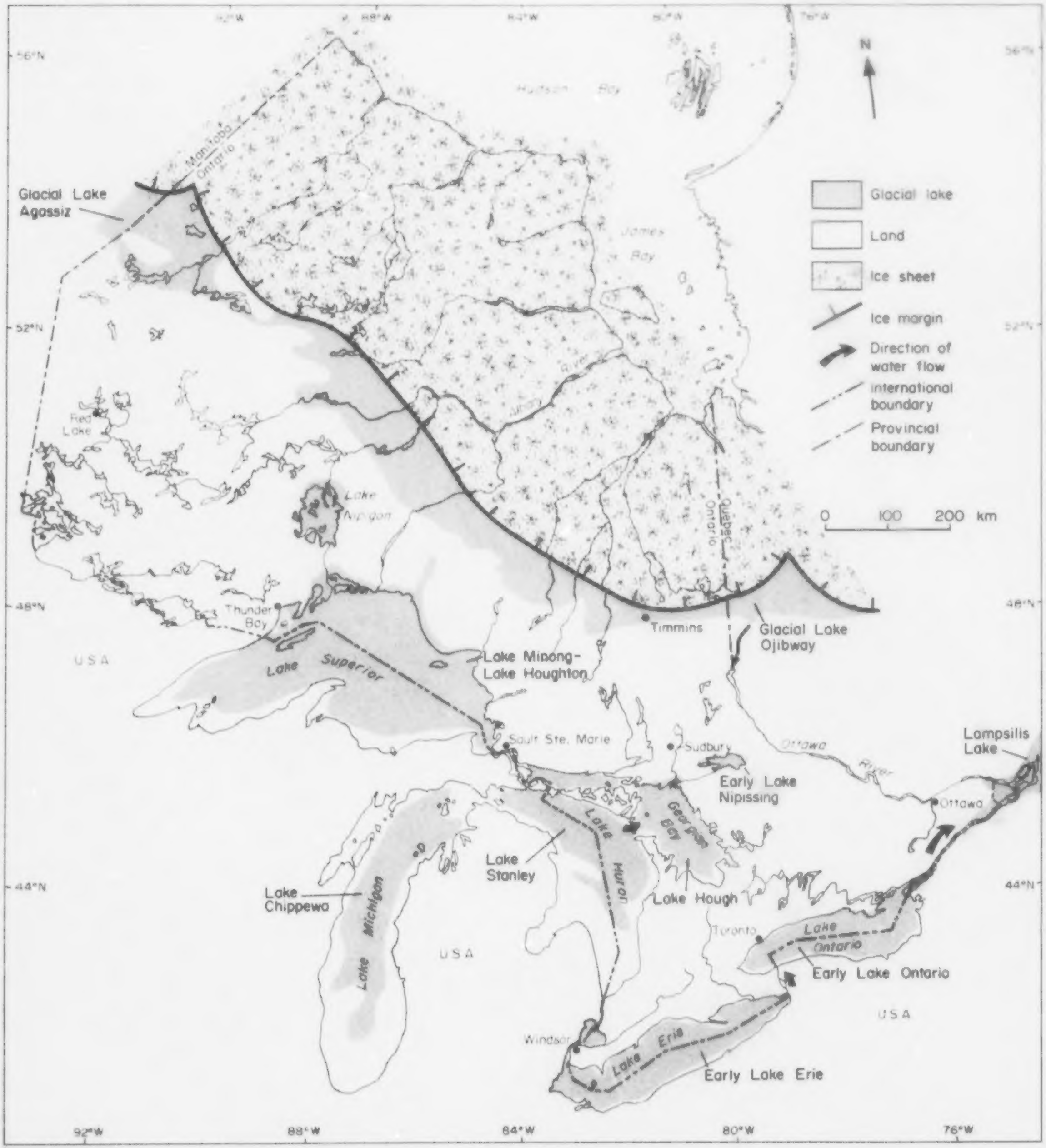
Hudson Bay
N
Land
Glacial lake
Glacial sea
Ice sheet
Ice margin
Direction of water flow
International boundary
Provincial boundary
Manitoba
Ontario
James Bay
Glacial Lake Agassiz
Red Lake
Lake Nipigon
Quebec
Ontario
0 100 200 km
Glacial Lake Kelvin
Thunder Bay
Glacial Lake Kaministikwia
Glacial Lake Duluth
Lake Superior
Timmins
Glacial Lake Sultan
Glacial Lake Ostrom
Glacial Lake Ogilvie
Glacial Lake Barlow
Sault Ste. Marie
Sudbury
Ottawa
Champlain Sea (falling level)
Glacial Lake Minong
USA
Lake Hough
Lake Stanley
Lake Huron
Lake Michigan
USA
Toronto
Lake Ontario
Early Lake Ontario
Lake Tonawanda
USA
Lake Chippewa
Windsor
Lake Erie
Early Lake Erie

h

**Figure 21.56h.** Deglaciation of Ontario: the beginning of the Holocene with the Marquette advance approximately 10 000 years ago. Many sources were used in the construction of the paleogeographic map presented, except where noted.

recharge areas. The common association of these deposits with less permeable till deposits that can lie beneath and/or on their flanks, makes ice-contact stratified drift deposits potentially valuable groundwater storage areas. The hummocky topography and kettle lakes commonly associated with ice-contact stratified drift deposits and landforms are very scenic. Therefore, the landforms have a good potential for recreational development.

Fine-grained sediments of glaciolacustrine and glaciomarine origin are generally poorly drained and usually have high water contents. Glaciomarine clays and silts pose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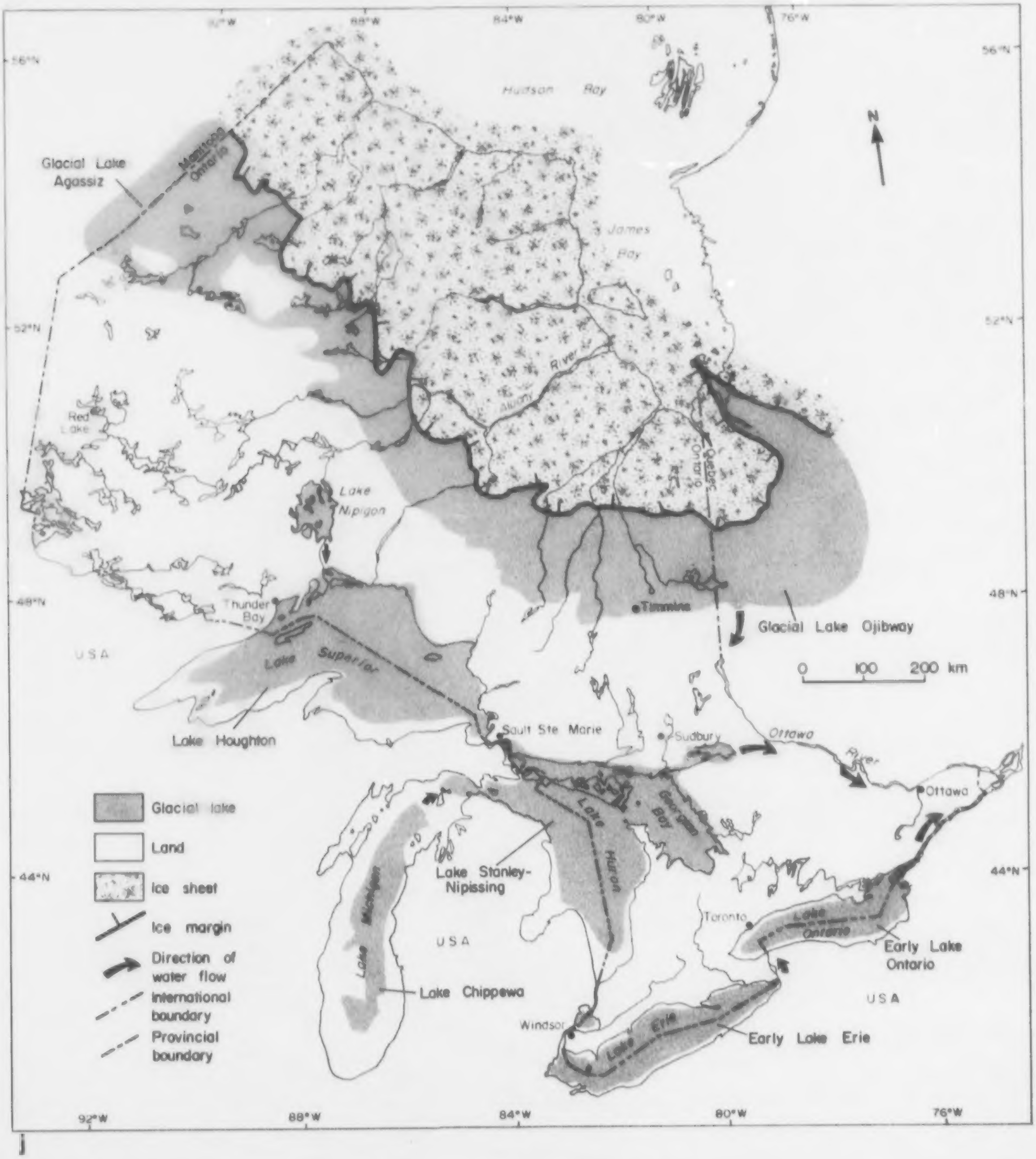
Ontario
Glacial Lake Agassiz
James Bay
Red Lake
Lake Nipigon
Thunder Bay
USA
Lake Superior
Lake Minong-Lake Houghton
Timmins
Glacial Lake Ojibway
Quebec
Ontario
Sault Ste. Marie
Sudbury
Ottawa River
Lampsilis Lake
Ottawa
Early Lake Nipissing
Georgian Bay
Lake Huron
Lake Stanley
Lake Chippewa
Lake Michigan
Lake Hough
Toronto
Lake Ontario
Early Lake Ontario
Windsor
Lake Erie
Early Lake Erie
Glacial lake
Land
Ice sheet
Ice margin
Direction of water flow
International boundary
Provincial boundary
0 100 200 km

**Figure 21.56l.** Deglaciation of Ontario: Ontario about 9000 years ago (Dyke and Prest 1987). Many sources were used in the construction of the paleogeographic map presented, except where noted.

particular problems to the engineer. Major landslides have occurred in the "sensitive clays" of the lower Ottawa and St. Lawrence river valleys and the Hudson Bay Lowland. The peculiar properties of these glaciomarine fine-grained sediments are summarized by Quigley (1980) and Torrance (1983). Mitchell and Klugman (1979) have reviewed several aspects of mass failures in "sensitive clays".

The most sensitive soils are the laminated sandy silt and silty clay rhythmites (G.A. Gorrell, Gorrell Resource Investigations, personal communications, 1987; Locat and Chagnon 1989) that were deposited as distal delta foresets or bottomsets in brackish waters of the Champlain Sea. The possible lateral connection to coarser foreset and topset beds provides the potential for the development of high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Red Lake
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Ontario
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Thunder Bay
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U.S.A.
Lake Superior
Lake Houghton
Sault Ste Marie
Sudbury
Ottawa River
Ottawa
Glacial lake
Land
Ice sheet
Ice margin
Direction of water flow
International boundary
Provincial boundary
Lake Huron
Georgian Bay
Lake Stanley-Nipissing
Lake Michigan
U.S.A.
Lake Chippewa
Toronto
Lake Ontario
Early Lake Ontario
U.S.A.
Windsor
Lake Erie
Early Lake Erie

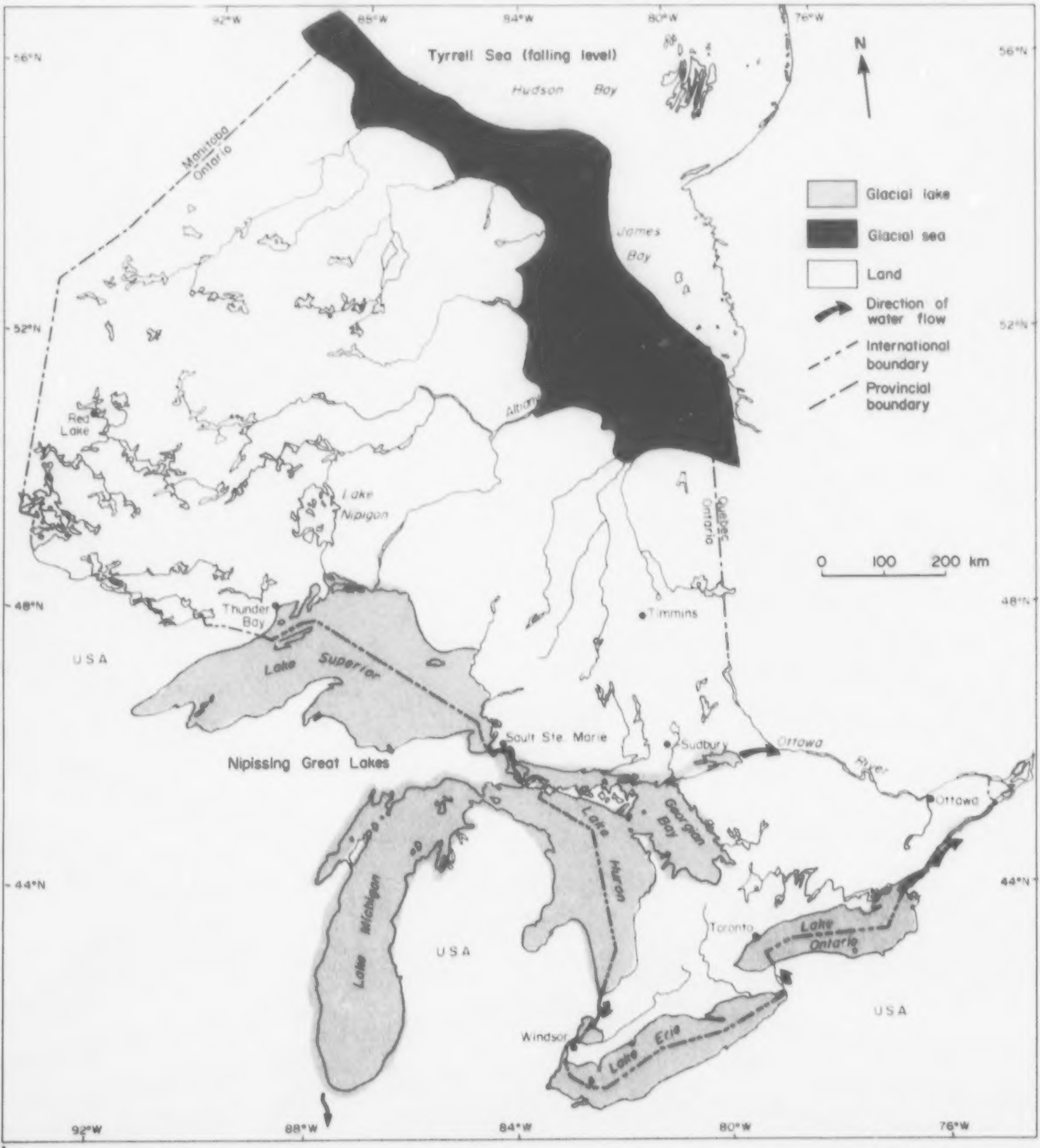
j

**Figure 21.56j.** Deglaciation of Ontario: approximately 8400 years ago during one of the Cochrane glacier advances (Dyke and Prest 1987). Many sources were used in the construction of the paleogeographic map presented, except where noted.

porewater pressures. This, combined with the metastable structure of the fine component of the couplet, and the ability of silt and fine sand to become quick, increases the sensitivity and the potential of large retrogressive failures in these sediments.

Thick organic deposits overlie the glaciomarine and marine sediments in the Hudson Bay Lowland. This thick blanket of organic material combined with the occurrence of permafrost conditions create difficulties for foundations.

Solid waste management is one of the most common environmental problems facing any community. A sanitary landfill is intended to isolate wastes from the environment and should be located and designed to keep soil, water, air and aesthetic pollution to a minimum (Mathewson 1981).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44°N
Tyrrell Sea (falling level)
Hudson Bay
N
Manitoba
Ontario
James Bay
Glacial lake
Glacial sea
Land
Direction of water flow
International boundary
Provincial boundary
Albany
Red Lake
Lake Nipigon
Quebec
Ontario
0 100 200 km
Timmins
Thunder Bay
USA
Lake Superior
Sault Ste. Marie
Nipissing Great Lakes
Sudbury
Ottawa River
Ottawa
Lake Huron
Georgian Bay
Lake Michigan
Toronto
Lake Ontario
USA
USA
Windsor
Lake Erie

k

**Figure 21.56k.** Deglaciation of Ontario: about 5000 years ago during the Nipissing Great Lakes (Dyke and Prest 1987). Many sources were used in the construction of the paleogeographic map presented, except where noted.

The geology of the site controls the hydrology. The permeability, structure, attitude and lithology of the hosting sediments are important factors to consider during site selection. In general, the lower the hydraulic conductivity (the degree of interconnection of the voids or pore spaces in a rock or soil that governs the movement of fluids through them), the more desirable the material. Fine-grained sediments generally have lower hydraulic conductivities than coarse-textured sediments. A material that may have a low conductivity, however, may contain structures such as fractures, faults, bedding planes or solution cavities, that will increase its hydraulic conductivity, making it undesirable for a landfill site (Mathewson 1981). Structures can be induced in sediments by desiccation, plant roots, sliding, subsidence and neotectonics. The attitude of bedding planes may provide avenues for leachate migration.

Dispersal rates of pollutants increase once the leachate enters the groundwater or surface water systems. Landfill sites should be located above the groundwater table in material of very low hydraulic conductivity. Engineering of geologically unsuitable sites is undesirable, but if necessary, should proceed with caution. Chemical diffusion through 1 m of clay can occur within 10 years (Johnson et al. 1989). Clay liners commonly constructed in geologically unsuitable landfill sites seldom exceed this thickness.

Knowledge of the properties of Quaternary sediments and their stratigraphy within shore bluffs and the foreshore area is essential for shore erosion, shore protection and management studies (Figure 21.58). The properties of Quaternary sediments affect: 1) the erodibility of the foreshore area and, therefore, toe erosion rates (Kamphuis 1986); 2) the types of processes that are effective in the degradation of the shore bluff; 3) the amount and flow paths of surface runoff; and 4) the distribution and flow paths of groundwater.

## CONSTRUCTION MATERIALS

The search for and evaluation of construction material, such as sand and gravel, are aided by an understanding of the Quaternary geology and history. Sand and gravel must be located close to where they will be used because large quantities are usually required. The bulk volume and low unit value of sand and gravel make transportation costs the most significant economic factor (Mathewson 1981).

Sand and gravel, a mineral aggregate, is used in concrete and asphaltic concrete. It provides the structural strength and reduces the quantity of cement needed. Clean, well-graded mineral aggregate that will form a strong bond with the cementing agent is desirable. Some mineral aggregates may react chemically with the cement, leading to failure of the concrete and possibly the structure (Mathewson 1981). Mineral aggregate is also needed without a cementing agent for road bases and wearing surfaces for roads and highways and foundation pad materials. This material should also be clean, well graded and able to be compacted to a high density. The presence of clay, silt or fine sand above specified limits may cause frost heave and ice lens problems (Mathewson 1981).

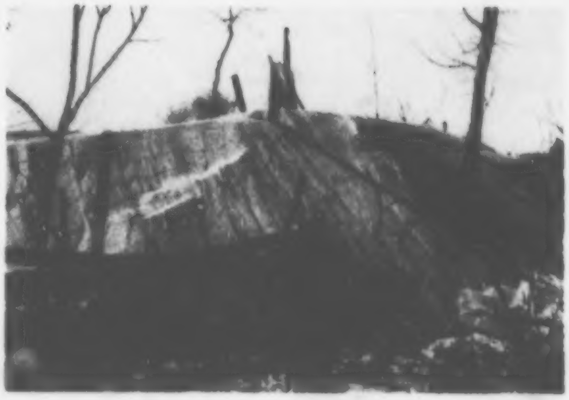

**Figure 21.57.** Windblown sand covering a fence, near Wesmeath, about 20 km east of Pembroke; the result of mismanagement of the land.

The geology of sand and gravel deposits and their environment of deposition are major factors that affect suitability for use as a construction material. Geological factors include: mineralogy; resistance to abrasion; soundness of particles; and the existence of coatings (Mathewson 1981). Environment of deposition factors include: controls on grain size; grading or sorting; and the presence or absence of fines.

The mineralogy can affect the suitability of a mineral aggregate deposit for its use in concrete. If the mineral aggregate deposit contains hydrated silicates, opaline silicates, chert, siliceous limestone, rhyolite or dacite clasts, an alkaline-aggregate reaction may occur which weakens the concrete (Mathewson 1981; Figure 21.59). Iron oxide, clay and calcite coatings on clasts may produce weak bonds between the aggregate particles and the cement, resulting in weak concrete. Mineral aggregate that is resistant to abrasion is used on roadways to produce and maintain the surface roughness required to produce skid-resistant pavement surfaces. Abrasion-resistant mineral aggregate may include quartz, crushed quartzite, granite and clay-free limestone and dolostone clasts (Mathewson 1981). Soundness is a measure of the weakness of individual aggregate particles. Fractures, discontinuities and weathering within individual aggregate clasts would lower the strength of the concrete if produced.

Quaternary sources of mineral aggregate include glaciofluvial ice-contact and outwash sediments, and glaciolacustrine and glaciomarine beach and delta deposits.

There is a high potential for finding mineral aggregates in ice-contact stratified drift deposits. However, abrupt changes in grain size within these features are not as yet very predictable, making long term extraction planning difficult. The linear nature of eskers also poses problems in planning, for a single esker may cross many property boundaries along its length (Cowan 1977).

Kames are not good targets for mineral aggregate extraction because of the moderate potential for coarse mineral aggregates, the usual small deposit size and the variability of the material (Cowan 1977). They may,

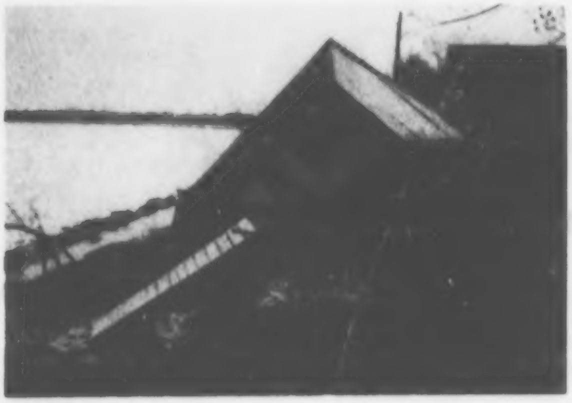

**Figure 21.58.** Damaged structure as a result of shoreline erosion, Port Burwell.

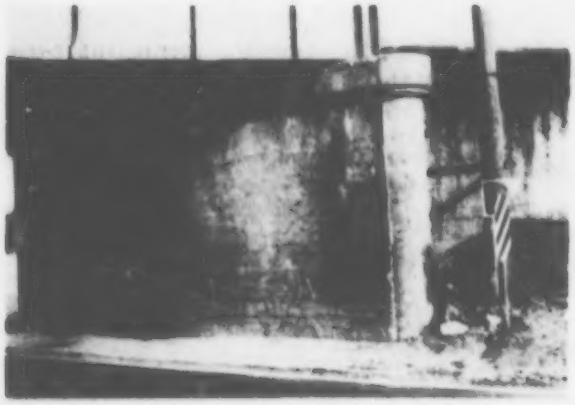

**Figure 21.59.** Concrete deterioration in a railway overpass near Sudbury, as a result of an alkaline-aggregate reaction. This reaction has created the polygonal cracking pattern in the concrete (photograph *courtesy of* G.R. Jones).

however, provide mineral aggregate material to meet local needs. Ice-marginal fans and deltas, and subaqueous fans are also affected by the same characteristics as kames, however, sediment distribution within these types of features is becoming better known (Smith and Ashley 1985; Gorrell, in press).

The larger ice-contact features such as interlobate moraines pose problems to both geologists and planners. The distribution of coarse mineral aggregates within these complex features is difficult to determine and the cost of exploration and development is often high (Cowan 1977).

Outwash deposits are ideal for mineral aggregates. They occur as sheets that can be thick and extensive. Their grain size is fairly uniform and large boulders are rare, except in the very proximal deposits. The probability of locating useful mineral aggregates in outwash deposits is high to moderate (Cowan 1977). The general consistency of outwash deposits makes long term extraction planning easier than in other mineral aggregate sources.

Abandoned beaches, spits and bars are also useful sources for mineral aggregates. Beach deposits are well stratified, and sorted into discrete size ranges. Deposits are uniform in grain size and oversized clasts are rare. The probability of finding coarse mineral aggregates is high where the beach has developed from a coarse substrate such as a stony till or gravelly ice-contact deposit (Cowan 1977). The narrow, linear nature of beach deposits combined with the tendency for them to be thin, usually less than 6 m, hinders their development as economic aggregate sources. One exception to this may be the thick deposits associated with baymouth bars. Delta deposits can be very large; however, the probability of finding coarse mineral aggregate material other than in the proximal zone is generally low (Cowan 1977).

Clay used for brick and tile is selected on the basis of its strength, colour, texture after firing and geochemical properties. Glaciolacustrine clay deposits have been used as raw material for the production of drainage tile and brick in the past, but, rarely today in Ontario (Guillet 1967, 1977).

Peat has many varied uses. Peat is used in agriculture for growing crops, as a soil conditioner and in greenhouses and gardens. It can be used as a fuel and as a filtration or absorption agent (Tarnocai 1989). Ontario has large deposits of peat; these are approximately 20% of the peatlands in Canada. Ontario peat should be suitable for both horticultural and fuel peat production. The use of peat for horticultural purposes is, however, far more promising (Monenco Ontario Limited 1981).

## Mineral Exploration

Deposits of the Laurentide Ice Sheet have been a hindrance to mineral exploration in the Canadian Shield for many years. In areas where the cover of Quaternary sediments is very thin and discontinuous, bedrock mapping and mineral exploration can proceed in traditional ways. In areas of more continuous drift cover, however, other methods must be used. This is particularly the case within the former basins of glacial lakes Barlow, Ojibway and Agassiz, where the cover of Quaternary sediments is very thick. In these areas, indirect mineral exploration methods such as geophysical and geochemical surveys and drift prospecting must be employed.

Classic geochemical techniques developed in non-glaciated areas do not work well in Ontario (*see* Fortescue, this volume). Residual soils are rare and, if present, are very thin and poorly developed. Most soils in Ontario are developed on transported parent materials, the products of glacial incorporation and transport with final deposition by ice, water or wind. The composition of the glacial drift and till in particular, however, can be very helpful in finding additional mineral resources.

Till is a primary product of glacial erosion and deposition. This, combined with its widespread distribution makes it an ideal medium to sample. The glacial debris of which till is composed represents the accumulation of the various rock types and sediments that the glacier has overridden. All components of till reflect the geology up-ice from the sample site. The concentration of any rock, mineral type or element in till decreases away from the source area (Figure 21.60). This decline in concentration approximates a negative exponential curve (Shilts 1976).

The body of till which is enriched in a certain rock, mineral or element is called a dispersal train or plume, and is often ribbon or fan shaped (DiLabio 1989). Lateral contacts of the dispersal train are often sharp, with concentrations of the enriched material dropping quickly to background levels. At the up-ice end or head of the dispersal train, concentrations are the highest. The tail of the dispersal train, or the exponential decline in concentration, is usually many times larger than the head and provides a much bigger target area than the head of the train or the original source. The main objective of a till geochemistry survey in mineral exploration is to simply detect the tail of a dispersal train, then follow it upglacier to its head, and then locate its source (Coker and DiLabio 1989).

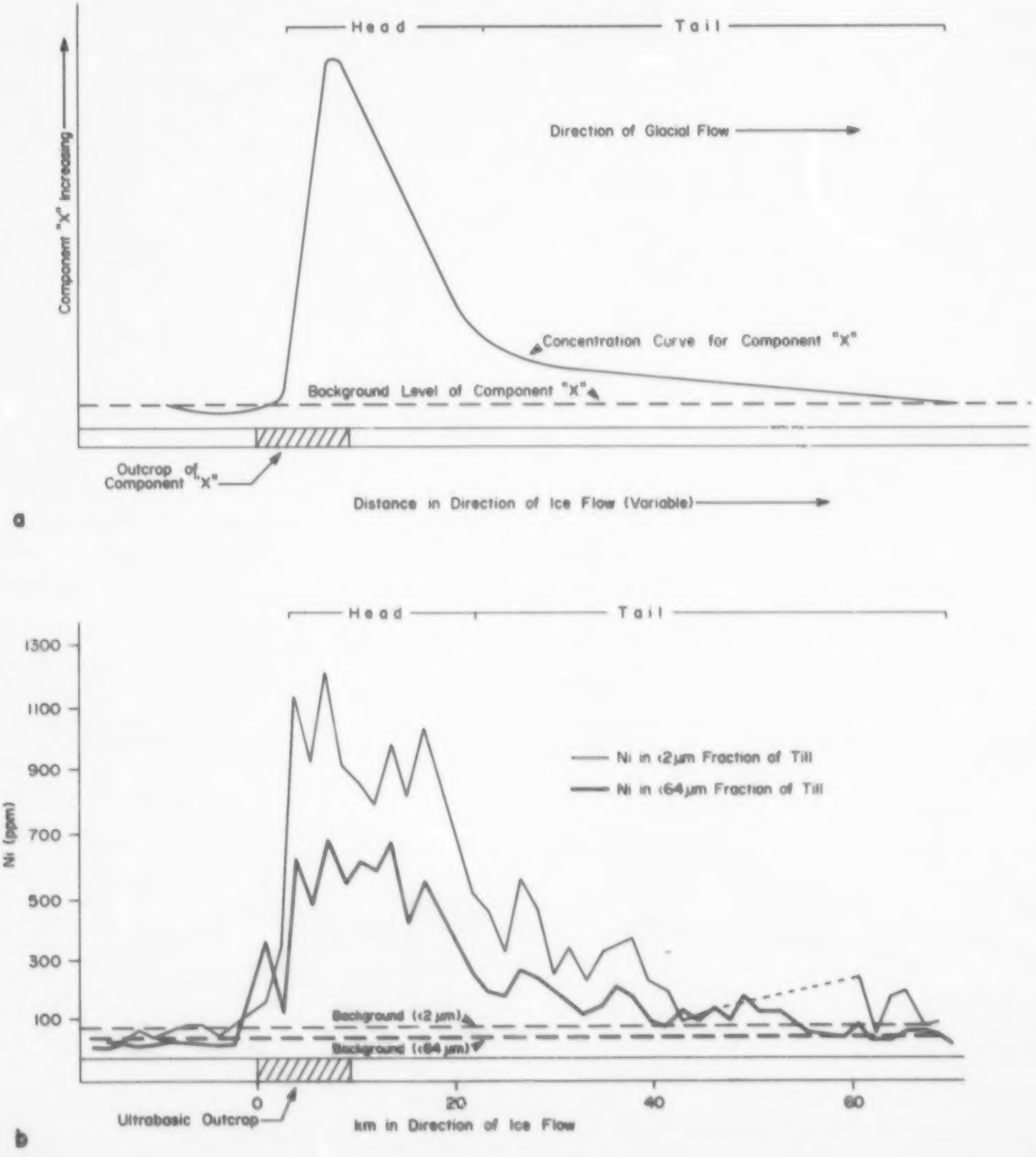

**Figure 21.60.** Component dispersal curves in till: a) idealized dispersal curve showing the head and tail of the common negative exponential curve; b) a regional dispersal curve of the nickel content of till samples from the Thetford Mines area, Quebec (*modified from* Shilts 1976).

Dispersal trains of distinctive boulders, minerals, major and/or trace elements have provided clues to the location of mineralized bedrock beneath the cover of Quaternary sediments (Shilts 1976, 1984a; Bolviken and Gleeson 1979; DiLabio 1989; Coker and DiLabio 1989). An idealized model of a glacial dispersal train that could apply to large parts of the Canadian Shield in Ontario, where only 1 till overlies an irregular bedrock surface, is illustrated in Figure 21.61. It should be noted that the position of the source in relation to topography and ice flow direction may alter the shape and extent of the dispersal train. In areas with significant local relief, the dispersal train may be trapped in valleys, broken into disjointed segments or even truncated (Coker and DiLabio 1989).

To detect dispersal trains derived from individual bedrock units, distinctive belts of rock, or mineralization in the preliminary stage of mineral exploration programs, sampling densities on the order of 1 till sample per square kilometre may suffice (Coker and DiLabio 1989). Sampling schemes designed to locate the head of dispersal trains are usually spaced on the order of 10s to 100s of metres (Coker and DiLabio 1989).

A fundamental knowledge of glacial deposits and their distribution, and the glacial history of the area, is essential for a successful drift prospecting program. Particular attention must be given to determining the glacial stratigraphy of the program area and the local and regional patterns of ice flow. Postglacial weathering and soil formation may result in the mobilization of elements in groundwater and soilwater. The enriched elements in the dispersal train may be concentrated in particular soil horizons or spread downslope, altering the original shape of the dispersal train.

In areas with a thick cover of glaciolacustrine sediments such as within the former basins of glacial lakes Barlow, Ojibway or Agassiz, sampling till for geochemical and lithological analysis requires trenching and drilling. Attention must be paid to the origin of the material encountered and sampled and to the stratigraphy, for it is often complex in glaciated basin areas. Within the basins of glacial lakes Barlow and Ojibway, a complex stratigraphy containing 3 or more till layers deposited by ice flowing in 3 different directions has been found beneath the surface glaciolacustrine deposits (Figure 21.62; Baker and Steele 1987; Baker, Steele and Fortescue 1984; Baker et al. 1985, 1986; Baker, Steele, McClenaghan and Fortescue 1984; Steele, Baker and McClenaghan 1986, 1989; Steele, McClenaghan and Baker 1986; Bird and Coker 1987; Veillette 1986). The bulk of glacial transport was achieved during the initial ice movement that deposited the lower part of the Matheson till, the silty sand surface till in the Kirkland Lake area (Veillette 1986; *see* Figure 21.62).

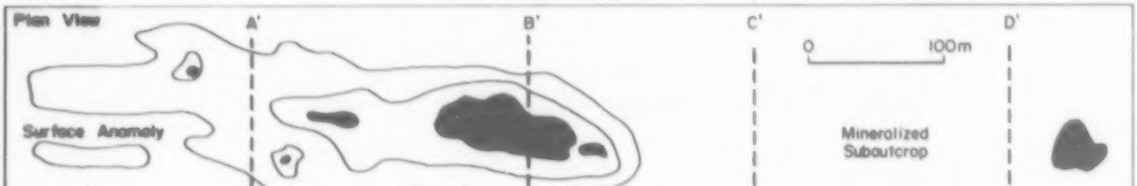

Figure 21.61. An idealized glacial dispersal model for areas of irregular bedrock topography (*modified from* Coker and DiLabio 1989; Miller 1984).

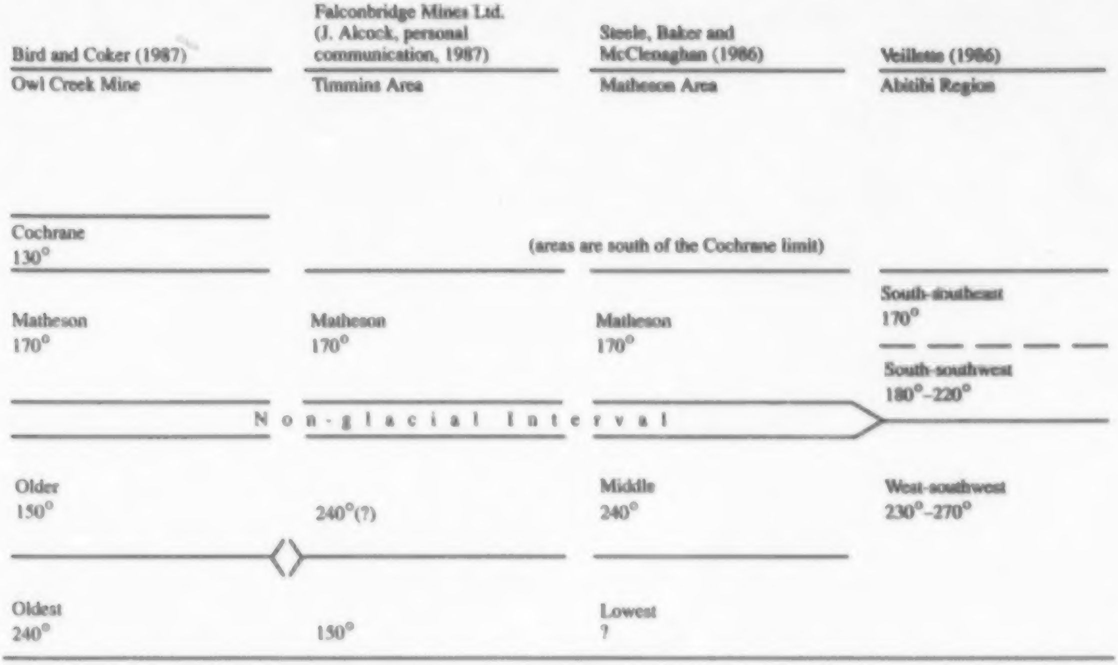
Bird and Coker (1987)
Owl Creek Mine
Falconbridge Mines Ltd. (J. Alcock, personal communication, 1987)
Timmins Area
Steele, Baker and McClenaghan (1986)
Matheson Area
Veillette (1986)
Abitibi Region
Cochrane
130°
(areas are south of the Cochrane limit)
Matheson
170°
Matheson
170°
Matheson
170°
South-southeast
170°
South-southwest
180°–220°
Non-glacial Interval
Older
150°
240°(?)
Middle
240°
West-southwest
230°–270°
Oldest
240°
150°
Lowest
?

**Figure 21.62.** Correlation of glacial stratigraphy—northeastern Ontario, northwestern Quebec (*modified from* Steele, Baker and McClenaghan 1986).

Multiple till layers occur in places beneath the surface glaciolacustrine sediments in the glacial Lake Agassiz basin. The stratigraphic relationships of tills in the Rainy River–Lake of the Woods area have been summarized and results of geochemical and lithological analyses prepared (Bajc 1991). Stratigraphic studies and mapping have also occurred in the Dryden area (Cowan 1987; Sharpe and Warman 1990; Minning et al., in press).

Mineral exploration is also hindered in areas of the Precambrian Shield where there is a thick cover of "calcareous" till. This till is composed of a high proportion of far-travelled material, carbonate rocks and minerals from the Hudson Bay Lowland, and therefore does not reflect local bedrock conditions (Geddes and Kristjansson 1986; Karrow and Geddes 1987; Dredge and Cowan 1989). The distribution and stratigraphic relationships of "carbonate" till, and the geochemistry of tills in the Hemlo and Beardmore–Geraldton areas of Ontario have been investigated (Geddes and Kristjansson 1986; Thorleifson and Kristjansson 1990).

The geochemistry of stream and lake sediments can also be investigated over large areas to characterize the geochemistry of the surface materials (Coker and Shilts 1979; Fortescue 1983, 1984; *see also* Fortescue, this volume). Geochemical anomalies found in these types of sediment samples, having been derived from the erosion of rock, till and other proglacial sediments, are very difficult to use to interpret source area; however, areas for follow up surveys can be identified. These types of surveys are also helpful in providing regional geochemical information for use in environmental and health related studies (*see* Fortescue, this volume).

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# Metallogeny

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## Chapter 22

# Metallogeny of Metallic Mineral Deposits in the Superior Province of Ontario

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## Abstract

**The time and space relationships between certain mineral deposit types and stages of the tectonic history are explored in this chapter. The tectonic history of the Superior Province within Ontario consists of 6 evolutionary stages, some of which are associated with the formation of metallic mineral deposits: 1) formation of greenstone assemblages in disparate tectonic settings; 2) accretion of greenstone assemblages into linear belts, on which are superimposed alkalic to calc-alkalic volcanic centres and batholith complexes; 3) erosion of emergent volcanic material to fill prismatic sedimentary basins; 4) progressive crustal-scale shortening, attendant underthrusting of greenstone subprovinces beneath sediment prisms, and related tectonic thickening; 5) localized melting of thickened crust; and 6) continued shortening by a combination of crustal underthrusting and lateral, subprovince-parallel slip.**

**Early volcanic assemblages formed between 3.03 and 2.80 Ga include platform and mafic-plain komatiitic assemblages, and spatially associated assemblages that contain minor felsic metavolcanic rocks and calc-alkalic plutons. These assemblages are preserved in parts of the Sachigo, Uchi, Wabigoon and Wawa subprovinces. Few known metallic mineral deposits occur in the older greenstone belt assemblages. Banded oxide-facies iron formation units occur within shallow-water platform or platform-foredeep successions. Komatiite-associated nickel-copper-platinum group sulphide deposits are not known within komatiite-bearing assemblages. No volcanic-associated, massive, base metal sulphide (VMS) deposits are known to occur. The presence of detrital gold and/or uranium deposits in metasedimentary rocks of the platform assemblages has been inferred, but not demonstrated. Parts of the northwestern Superior Province may have experienced a collisional-type orogeny, which terminated at about 2.9 Ga. At least one gold deposit may have formed during this orogeny.**

**A second period of greenstone belt construction took place between approximately 2.75 and 2.70 Ga. Diverse volcanic and sedimentary assemblages formed in disparate tectonic settings in the ancestral Uchi, Wabigoon, Wawa and Abitibi subprovinces. Volcanic assemblages may have formed in mafic plain, arc, rifted arc or back-arc tectonic settings. Intragreenstone belt turbidite successions formed late during the construction of the greenstone belts. Metallic mineral deposits that occur within, and were coeval with, these assemblages and cogenetic intrusions include: 1) volcanic-associated, massive, base metal sulphide (VMS) mineralization; 2) komatiite-associated nickel-copper-platinum group mineralization; 3) banded iron formation; 4) porphyry intrusion-related copper-molybdenum; and 5) nickel-copper-platinum group mineralization in synvolcanic mafic and ultramafic intrusions. All unequivocal VMS deposits formed between 2.75 and 2.7 Ga. Most VMS deposits occur in successions that contain chemically evolved rhyolitic (high-silica rhyolite) and andesitic (icelandite) rocks. These deposits, and some of the spatially and temporally associated rocks, formed during an extensional period of arc, rifted arc, or back-arc development. All known komatiite-associated nickel-copper-platinum group deposits occur in the Abitibi Subprovince, within assemblages that are younger than 2.74 billion years old. These deposits formed in spreading ridge, rifted back-arc, or rifted arc settings. The Thierry deposit, in the Uchi Subprovince, and the Shebandowan deposit, in the Wawa Subprovince, are associated with deformed ultramafic and mafic intrusions, which may have a komatiitic affinity. Few porphyry copper-molybdenum deposits occur in the Superior Province of Ontario. Some occur within subvolcanic, foliated to massive, granodiorite intrusions that were emplaced into a collisional tectonic setting during the calc-alkalic constructional phase of the volcano-plutonic arcs. Nickel-copper-platinum group and chromite mineralization occur within subvolcanic, mafic to ultramafic, komatiitic and tholeiitic intrusions that were emplaced into komatiite- and tholeiite-bearing assemblages during the construction of the volcano-plutonic arcs. The intrusions may represent high-level magma chambers that formed at a spreading ridge, in a back-arc basin, or during the primitive stages of arc construction. Banded iron formation units also accumulated in the younger (2.8 to 2.7 Ga) volcanic assemblages. Iron formation units associated with accretionary sedimentary prisms formed following the construction of the volcano-plutonic arcs, during a period of greenstone belt uplift and erosion at about 2.7 Ga.**

**The assembly of the disparate volcanic assemblages into volcanic-plutonic subprovinces was temporally and spatially associated with the formation of alkalic intrusion-associated copper-gold deposits, a second stage of porphyry intrusion-related copper-molybdenum deposits, and mafic to ultramafic intrusion-associated nickel-copper-platinum group deposits. Some gold deposits in the central part of the Uchi greenstone belt may also have formed during the construction of subduction-related volcano-plutonic arcs, or during the assembly and juxtaposition of allochthonous assemblages, but prior to the regional folding and faulting related to the accretion of the Superior Province during the Kenoran Orogeny. During the final stages of arc coalescence, between 2.7 and 2.69 Ga, the volcanic-plutonic subprovinces were uplifted due to tectonic thickening, and were eroded to yield detritus that filled prismatic sedimentary basins. No metallic mineral deposits formed during this regional-scale emergence.**

**Late unconformable basins within the assembled volcano-plutonic subprovince were the locus of Timiskaming-like assemblages, consisting of alluvial-fluvial sedimentary rocks interlayered with minor**

**amounts of alkalic metavolcanic rocks. The development of these Timiskaming-like assemblages between 2.70 and 2.65 Ga may have corresponded with a change in regional-scale tectonic style within the volcano-plutonic arcs from dominantly collisional to dominantly extensional. Poorly described and preserved iron formation units formed within Timiskaming-like metasedimentary assemblages, perhaps in an extensional setting. Alkalic, intrusion-associated, copper-gold deposits in the Abitibi and Wawa subprovinces are spatially associated with regional deformation zones and Timiskaming-like assemblages.**

**Between 2720 and 2660 Ma, the Superior Province formed by the southward accretion of volcanic-plutonic subprovinces (Uchi, Wabigoon, Wawa and Abitibi) and adjacent sedimentary prisms (English River and Quetico) during the Kenoran Orogeny. Attendant metamorphism and partial melting of the tectonically thickened crust yielded peraluminous and calc-alkalic tonalitic magmas, which were emplaced along subprovince boundaries, adjustment structures, and, less commonly, within subprovinces. Some of the peraluminous magmas underwent extensive chemical fractionation to yield rare element-bearing pegmatites. Late mafic intrusions were also emplaced along, or near, regional deformation zones during the Kenoran Orogeny and are host to nickel-copper-platinum group mineralization. Continued shortening across the Superior Province was accommodated, in part, by a combination of underthrusting and lateral, subprovince-parallel slip. Many gold deposits formed late, during or following this tectonic stage in, and adjacent to, regional deformation zones.**

# INTRODUCTION

The Archean Superior Province represents an area of nearly 2 million square kilometres, consisting of small areas of *circa* 3.0 to 3.2 billion-year-old crust, and large areas of 2.8 to 2.7 billion-year-old crust (Card and Ciesielski 1986). It consists of east- to east-northeast-trending subprovinces which display different rock assemblages, ages, metamorphic histories, structural styles and metallogenic histories. The Superior Province is truncated on the west by the Trans–Hudson Orogen and on the east by the Grenville and New Quebec (formerly Labrador Trough) orogens (Hoffman 1989). Other areas of Proterozoic deformation within, or at the margins of, the Superior Province include the Midcontinent Rift and the Kapuskasing Structural Zone. The western, northern and southeastern boundaries of the Superior Province with the Trans–Hudson and Grenville orogens are tectonic and the loci of thrusting and transcurrent faulting (Card 1990). The southern and eastern contacts with the Penokean and New Quebec orogens represent Proterozoic supracrustal successions that unconformably overlie, and were thrust upon, the Superior Province (Card 1990).

The Superior Province is host to a broad spectrum of metallic mineral deposit types (Franklin and Thorpe 1982). The types and distributions of metallic deposits in various Archean terranes, including the Superior Province, have been described by several workers (Boyle 1976; Gross 1977; Hutchinson 1981; Ayres and Cerny 1982; Franklin and Thorpe 1982; Groves and Batt 1984; Breaks, Cherry et al. 1985; Thurston and Chivers 1990). Many of these discussions dwell on the type of metallic mineral deposits present and spatial relationships between deposit types and associated rocks.

In this chapter, some metallic deposit types in the Superior Province of Ontario are described. Utilizing a magmatic, metamorphic and tectonic framework for the Superior Province in Ontario (Thurston et al., this volume; Williams, Stott, Thurston et al., this volume), the time and space distributions of the deposit types and their spatially and temporally associated tectonic elements are illustrated to develop a metallogenic history of the Superior Province in Ontario. Metallic mineral deposit types discussed in this chapter are selected on the basis of their economic importance, and to illustrate key spatial and temporal relationships between deposit types and tectonic elements. For comparative purposes, some deposits from Quebec are also referenced.

Metallic mineral deposit types addressed include: 1) volcanic-associated, massive, base metal sulphide (VMS); 2) komatiite-associated nickel-copper-platinum group element; 3) iron formation; 4) mafic-ultramafic intrusion-hosted nickel-copper-platinum group element and chromite; 5) rare-metal and uraniferous pegmatites; 6) porphyry copper-molybdenum and copper-molybdenum-gold; and 7) gold. There are other deposit types present in the Superior Province (Thurston and Chivers 1990), such as the Crane tungsten-quartz vein deposit in the Mishibishu Lake area (Bennett and Thurston 1977), but they are not fully documented and are generally of lesser importance and are not considered here.

# OVERVIEW OF THE SUPERIOR PROVINCE

The Superior Province is subdivided into 4 subprovince types, based on differing rock assemblages, and structural, metamorphic and metallogenic histories (Card and Ciesielski 1986; Card 1990): 1) volcanic-plutonic or "granite-greenstone" subprovinces, which represent various depositional settings (summarized in Williams, Stott, Thurston et al., this volume); 2) metasedimentary subprovinces, which resemble accretionary prisms; 3) plutonic subprovinces, which may represent deeper erosional levels of volcanic-plutonic terranes (Beakhouse, this volume; Thurston et al., this volume); and 4) high-grade gneiss subprovinces, most of which are deeper erosional levels of the other subprovince types (Williams, Stott, Thurston et al., this volume).

The granite-greenstone subprovinces consist of low-grade, metavolcanic-metasedimentary successions intruded by granitoid rocks (Card and Ciesielski 1986). These subprovinces contain the most diverse types of metallic mineral deposits, including: 1) VMS; 2) iron formation; 3) komatiite-associated nickel-copper-platinum group element sulphide; 4) magmatic nickel-copper-platinum group elements and chromite in mafic and ultramafic intrusions; 5) porphyry copper-molybdenum and porphyry copper-gold-molybdenum; and 6) gold.

The metasedimentary subprovinces, which separate some volcano-plutonic subprovinces, consist mainly of deformed and metamorphosed turbidites and associated granitoid intrusions including anatectic, two-mica granites (Card and Ciesielski 1986). Rare element and uraniferous pegmatites are characteristic metallic deposit types in these subprovinces.

High-grade gneiss belts are characterized by upper amphibolite- and granulite-facies gneiss of supracrustal and plutonic origin (Card and Ciesielski 1986). Rare element pegmatites represent the most economically significant deposit type in these subprovinces, although more varied mineral deposit types occur in other Archean high-grade terranes (Katz 1988).

Subprovince boundaries are generally marked by abrupt changes in rock type and structural and metamorphic styles (Thurston and Breaks 1978) and hence, magnetic signature (Hall and Brisbin 1982). Generally, the subprovince boundaries are loci of shearing and plutonism (Percival and Williams 1989). Some subprovince boundaries have associated gold mineralization (Colvine 1989; Card et al. 1989; Card 1990).

## METALLIC MINERAL DEPOSITS FORMED DURING EARLY TECTONIC AND MAGMATIC EVENTS IN VOLCANIC-PLUTONIC SUBPROVINCES

The greatest diversity of metallic mineral deposits within the Superior Province in Ontario occurs within the granite-greenstone subprovinces. Deposits consist of 2 types: 1) those related to tectonic setting of rock assemblages; and 2) those related to orogenic processes, and hence are superimposed upon rock assemblages (e.g., gold). Examples of type 1 deposits include VMS and magmatic, intrusion-hosted nickel-copper-platinum group element mineralization. Examples of type 2 deposits include gold and rare element pegmatites.

### Volcanic-Associated, Massive, Base Metal Sulphide Mineralization (VMS)

Archean VMS deposits belong to a general class of stratabound sulphide deposits similar to those in younger orogens (Sangster 1972; Franklin et al. 1981; Franklin and Thorpe 1982; Lydon 1984; Franklin 1986). The Archean deposits are classified into copper-zinc and zinc-copper-(lead) on the basis of metal ratios in the deposit, the degree of water-rock interaction (as seen in the types of volcanic products in the deposit area), and the size and type of alteration systems (Gilmour 1976; Morton and Franklin 1987).

The idealized deposit, typified by many in the Noranda camp, consists of a concordant mineralized lens containing greater than 60% massive sulphide minerals (Figure 22.1; Sangster 1972; Sangster and Scott 1976; Lydon 1984). The upper part of the massive sulphide zone is generally sphalerite-, pyrite- and galena-rich, whereas the base is generally chalcopyrite-rich (Franklin et al. 1981). The massive sulphide lens is stratigraphically underlain by a discordant, chalcopyrite-rich stringer zone, contained in a pipe of hydrothermally altered rock (Figure 22.2; Sangster 1972; Sangster and Scott 1976; Lydon 1984). The upper contact of the massive sulphide lens is usually sharp, but the lower contact is usually gradational into a stringer zone (Sangster 1972). The stringer zone is interpreted as the locus of hydrothermal fluid discharge (Sangster 1972; Lydon 1984). Some deposits do not display this complete spectrum of mineralization habits (Morton and Franklin 1987). For example, deposits in the Sturgeon Lake camp have large stratiform zones of alteration rather than spatially restricted alteration pipes and are not copper-rich (Morton and Franklin 1987).

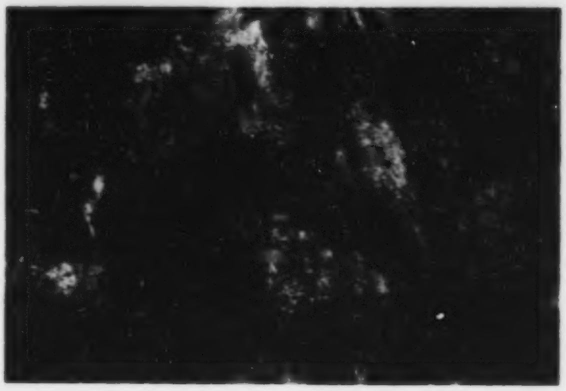

**Figure 22.1.** Volcanic-associated, massive, base metal sulphide (VMS) mineralization from the Kidd Creek Mine (Falconbridge Limited), Timmins. Banding is defined by alternating layers of sphalerite and pyrite.

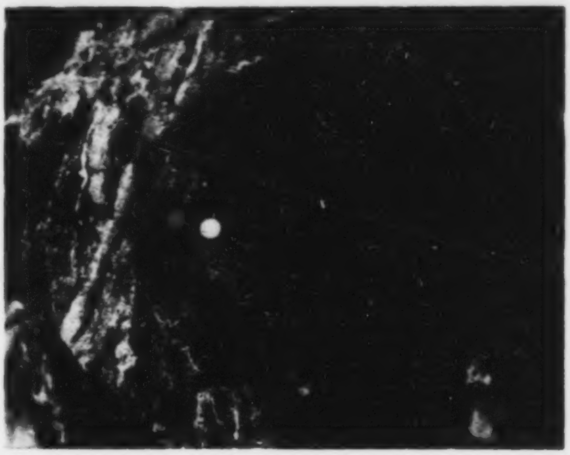

**Figure 22.2.** Pyrite and chalcopyrite stockwork veins that cut chloritized volcanic rock beneath the Kam Kotia deposit, Timmins. This style of alteration and vein mineralization occurs within pipe-shaped zones beneath many volcanic-associated, massive, base metal sulphide (VMS) deposits.

## REGIONAL DISTRIBUTION OF VMS DEPOSITS

VMS deposits occur in most subprovinces, but are most abundant in the Abitibi greenstone belt, Abitibi Subprovince (Figure 22.3; Table 22.1). The principal VMS camps in the Abitibi greenstone belt include those of the Kidd Creek, Kamiskotia, Noranda, Mattagami and Chibougamau areas (*see* Figure 22.3). Principal deposits in other subprovinces include the Manitouwadge and Winston Lake camps in the Wawa Subprovince, the Sturgeon Lake camp in the western Wabigoon Subprovince, and the South Bay camp, in the Uchi Subprovince (*see* Figure 22.3; Table 22.1).

## CAMP- TO DEPOSIT-SCALE ROCK ASSOCIATIONS

Archean felsic metavolcanic rocks are subdivided into FI, FII and FIII types based on trace element geochemistry (Condie 1976, 1981). Chemically distinctive rhyolites (Table 22.2), termed FIII-rhyolites (Condie 1981; Lesher et al. 1986), occur in VMS-bearing successions in the Superior Province, including the Noranda (Ujike and Goodwin 1987), Mattagami Lake, Kidd Creek and Kamiskotia camps (Hart 1984; Lesher et al. 1986; Barrie 1990) in the Abitibi greenstone belt, and at the South Bay Mine (Thurston 1981; Lesher et al. 1986) in the Uchi Subprovince. Type FIII is subdivided into FIIIa and FIIIb (*see* Table 22.2; Lesher et al. 1986). Type FIIIa exhibits moderately negative Eu anomalies, low Zr:Y ratios (4 to 7), high abundances of Sc, and intermediate abundances of high field strength elements (Lesher et al. 1986). Type FIIIb exhibits a pronounced negative Eu anomaly, low Zr:Y ratios (2 to 6), high abundances of high field strength elements, and low abundances of Sc and Sr (Lesher et al. 1986). Most VMS deposits in the Superior Province are spatially associated with type FIII felsic metavolcanic rocks (Lesher et al. 1986).

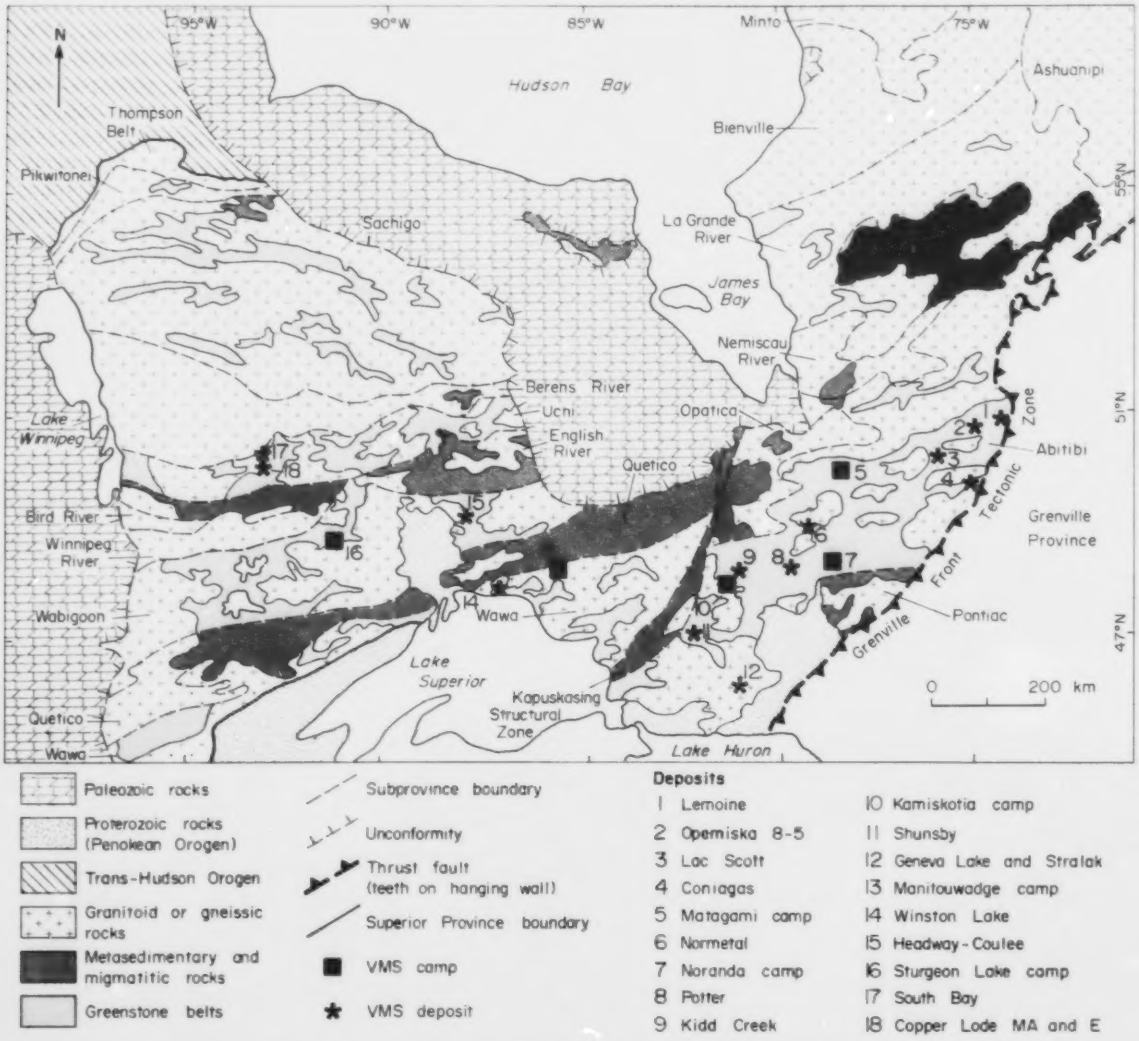
95°W
90°W
85°W
Minto
75°W
N
Hudson Bay
Ashuanipi
Thompson Belt
Bienville
Pikwitonei
55°N
Sachigo
La Grande River
James Bay
Nemiscau River
Berens River
Uchi
Opatica
Lake Winnipeg
English River
Quetico
Zone
51°N
Abitibi
Tectonic
Bird River
Grenville Province
Winnipeg River
Front
Wabigoon
Wawa
Pontiac
Lake Superior
Grenville
47°N
Kapuskasing Structural Zone
0
200 km
Quetico
Wawa
Lake Huron
1
2
3
4
5
6
7
8
9
10
11
12
13
14
15
16
17
18
Paleozoic rocks
Proterozoic rocks (Penokean Orogen)
Trans-Hudson Orogen
Granitoid or gneissic rocks
Metasedimentary and migmatitic rocks
Greenstone belts
Subprovince boundary
Unconformity
Thrust fault (teeth on hanging wall)
Superior Province boundary
VMS camp
VMS deposit
Deposits
1 Lemoine
2 Opemiska 8-5
3 Lac Scott
4 Coniagas
5 Matagami camp
6 Normetal
7 Noranda camp
8 Potter
9 Kidd Creek
10 Kamiskotia camp
11 Shunsby
12 Geneva Lake and Stralak
13 Manitouwadge camp
14 Winston Lake
15 Headway-Coulee
16 Sturgeon Lake camp
17 South Bay
18 Copper Lode MA and E

**Figure 22.3.** Distribution of volcanic-associated, massive, base metal sulphide (VMS) deposits in the Superior Province.

Andesite flows and sills, rich in $P_2O_5$, called icelandites (Macdonald et al. 1990), are spatially associated with the FIII-rhyolites in the Kamiskotia and Kidd–Munro assemblages in the southern Abitibi greenstone belt (Barrie 1990; Binney and Barrie 1990; Brisbin et al. 1990). Insufficient data are available to determine the general distribution of these distinctive rhyolites and andesites in other greenstone belts of the Superior Province, although the rhyolites do not appear to be present in the western Wabigoon Subprovince (Lesher et al. 1986). The basalt–icelandite–FIII-rhyolite suite was derived by partial melting of primitive mantle and low-pressure, plagioclase-dominated fractional crystallization in a high-level, subvolcanic, zoned mafic-felsic magma chamber (Lesher et al. 1986; Ujike and Goodwin 1987; Barrie 1990; Macdonald et al. 1990).

The VMS deposits in the Sturgeon Lake area are spatially associated with less chemically evolved FII-rhyolites (*see* Table 22.2; Lesher et al. 1986). Whereas the association of VMS deposits with FII-rhyolites is common in Phanerozoic terranes, it is unusual in the Superior Province, where the spatial association with FIII-rhyolites predominates (Lesher et al. 1986).

## VOLCANIC-ASSOCIATED MASSIVE BASE METAL-BEARING VOLCANIC ASSEMBLAGES

VMS deposits in the Superior Province of Ontario occur in the subaqueously deposited parts of 2 different assemblage types (*see* Table 22.1): 1) bimodal mafic-felsic or intermediate-felsic effusive volcanic successions; and 2) in felsic metavolcanic rocks forming part of komatiite-tholeiite successions . At a deposit scale, footwall rocks are also variable, and the proportion of rock types may differ from that representative of the assemblage as a whole. Intermediate to felsic footwall rocks occur beneath many, but not all, deposits (*see* Table 22.1; Thurston 1981; Trowell 1986; Ujike and Goodwin 1987; Brisbin et al. 1990; Binney and Barrie 1990). Mafic volcanic footwall rocks occur beneath some deposits in the Noranda and Kamiskotia camps; and the Potter deposit, in the Abitibi greenstone belt (*see* Table 22.1).

### Bimodal Assemblages

Many, but not all, VMS deposits in the Superior Province occur in bimodal successions of intermediate to felsic flows and pyroclastic rocks (*see* Table 22.1; Dimroth et al. 1983; Thurston 1986; Morton and Franklin 1987; Gibson and Watkinson 1990). These bimodal volcanic successions represent stratovolcanoes (Thurston et al. 1978) or central volcanic complexes (Morton 1984; Dimroth et al. 1985). Some of these volcanic successions are alternatively interpreted as large composite cones (Goodwin 1979; Dimroth et al. 1985; Thurston 1986; Gibson and Watkinson 1990). Some VMS deposits, such as those in the Sturgeon Lake area, occur in volcanic successions dominated by cauldron-related fragmental rocks (Morton et al. 1990) that were erupted and deposited under relatively shallow water conditions (Morton and Franklin 1987). Other VMS deposits, such as some in the Noranda camp, are associated principally with cauldron-related, effusive metavolcanic rocks (Spence and de Rosen-Spence 1975; Dimroth et al. 1983; Morton and Franklin 1987; Gibson and Watkinson 1990) that were erupted in deeper water settings (Morton and Franklin 1987).

Within the volcanoes, some VMS deposits, such as those in the Confederation Lake, Sturgeon Lake and Noranda areas, occur within the middle and upper metavolcanic units that were erupted during a period of cauldron subsidence and subsequent resurgence (Thurston 1986; Morton 1984; Gibson and Watkinson 1990). Some volcanic vents were localized along synvolcanic faults, which served as a principal control on the localization of individual VMS deposits (Knuckey et al. 1982; Gibson and Watkins 1990).

### Komatiite-Tholeiite Assemblages

This assemblage type consists of diverse rock types, including komatiites, tholeiitic basalts, icelandite, FIII-rhyolite, calc-alkalic andesite and rhyolite, and related intrusions (Jackson and Fyon, this volume). Not all rock types need be present, and further subdivisions of this assemblage type are based on the presence or absence of komatiite (Jackson and Fyon, this volume). To emphasize this diversity, the settings and rock associations of 2 VMS deposits in the Abitibi greenstone belt, the Potter and Kidd Creek mines, are discussed in detail.

#### KOMATIITE-THOLEIITE-DOMINATED SUCCESSIONS

In the eastern part of the Kidd–Munro assemblage, southern Abitibi greenstone belt, the Potter copper-zinc VMS deposit (Coad 1976; Green and MacEachern 1990) occurs within a succession dominated by komatiite and tholeiite (Jackson and Fyon, this volume). Spinifex-textured and massive komatiites occur within a paleo-lava lake, which is overlain by mafic, tholeiitic, fragmental rock, thin beds of dacitic ash, and rare chert (Coad 1976; Arndt 1986). This komatiite unit is on strike with the Pyke Hill komatiite flows to the east (Arndt 1986). The basaltic fragmental unit consists of scoria, agglomerate and welded spatter (Coad 1976). To the east, the volcanic rocks are intruded by, or are in faulted contact with, gabbros from the upper portion of the Centre Hill layered mafic-ultramafic intrusion (MacRae 1969). The intrusive relationship between the Centre Hill Complex and the adjacent basaltic fragmental unit, and their similar chemical composition suggest that the Centre Hill Complex was consanguineous with the mafic fragmental rock (Coad 1976).

Massive sulphide mineralization at the Potter Mine was localized over the faulted contact between the Centre Hill Complex and the komatiite lava lake (MacRae 1969; Coad 1976). The fragmental rock thickens along strike across the fault, and neither the fragmental rock nor the mineralized horizons are offset on either side of the fault (Coad 1976). Chlorite alteration and disseminated sulphide mineralization occur within, and adjacent to, the fault (Coad 1976).

**Table 22.1** General characteristics of Archean volcanic-associated, massive, base metal sulphide deposits in the Superior Province of Ontario.

| Name (Reference)[1] | Subprovince | Greenstone Belt | Assemblage Name | Assemblage Type | Rock Association |
|---|---|---|---|---|---|
| Stralak (1,9) | Abitibi | Benny | Geneva | bimodal | basalt, andesite, dacite, rhyolite |
| Lake Geneva (1,9) | Abitibi | Benny | Geneva | bimodal | basalt, andesite, dacite, rhyolite |
| Kam Kotia (1,3) | Abitibi | Abitibi | Kamiskotia | bimodal | mafic volcanic rocks, icelandite, FIII-rhyolite, layered tholeiitic intrusion |
| Jameland (1,3) | Abitibi | Abitibi | Kamiskotia | bimodal | icelandite, FIII-rhyolite, layered tholeiitic intrusion |
| Canadian Jamieson (1,3) | Abitibi | Abitibi | Kamiskotia | bimodal | iron-rich basalt, icelandite, FIII-rhyolite, layered tholeiitic intrusion |
| Genex (3) | Abitibi | Abitibi | Kamiskotia | bimodal | icelandite, FIII-rhyolite, layered tholeiitic intrusion, iron-rich tholeiitic basalt |
| Potter (1) | Abitibi | Abitibi | Kidd–Munro | komatiite-tholeiite | basalt, komatiite, layered tholeiitic intrusion |
| Kidd Creek (1,2) | Abitibi | Abitibi | Kidd–Munro | komatiite-tholeiite | basalt, icelandite, FIII-rhyolite, carbonaceous argillite, ultramafic rocks |
| Winston Lake (4) | Wawa | Schreiber – Terrace Bay | Schreiber | bimodal | felsic flows and pyroclastic rocks, sediments, iron- and magnesium-rich tholeiitic basalt, layered tholeiitic intrusion |
| Willroy (1,6,7) | Wawa | Manitouwadge – Hornepayne | unnamed | bimodal? | basalt, sericite schist (felsic volcanic?), sedimentary rocks, diorite, tonalite-granodiorite |
| Big Nama Creek (1,6,7) | Wawa | Manitouwadge – Hornepayne | unnamed | bimodal? | sericite schist (felsic volcanic?), pegmatite, basalt |
| Willecho (1,6,7) | Wawa | Manitouwadge – Hornepayne | unnamed | bimodal? | iron formation, mafic volcanic rocks, sillimanite and garnet schists |
| Geco (1,6,7) | Wawa | Manitouwadge – Hornepayne | unnamed | bimodal? | iron formation, sedimentary rocks, sericite schist (felsic volcanic?), mafic volcanic rocks, pegmatite, |
| Headway–Coulee area (1,8) | Wabigoon | Onaman–Tashota | unnamed | tholeiite? | basalt, minor felsic ash and tuffs |
| Mattabi (1,5) | Wabigoon | Sturgeon Lake | Central Sturgeon | bimodal, cauldron complex | basalt, rhyolite, intermediate to felsic pyroclastic rocks, breccia, debris flows, felsic domes and flows, trondhjemite-granodiorite |
| F–Group (5) | Wabigoon | Sturgeon Lake | Central Sturgeon | bimodal, cauldron complex | basalt, rhyolite, intermediate to felsic pyroclastic rocks, breccia, debris flows, felsic domes and flows, trondhjemite-granodiorite |
| Sturgeon Lake (1,5) | Wabigoon | Sturgeon Lake | Central Sturgeon | bimodal, cauldron complex | basalt, rhyolite, intermediate to felsic pyroclastics rocks, breccia, debris flows, felsic domes and flows, trondhjemite-granodiorite |
| Lyon Lake and Creek Zone (1,5) | Wabigoon | Sturgeon Lake | Central Sturgeon | bimodal, cauldron complex | fragmental rhyolite, dacite flows, mafic volcanic rocks, iron formation, wacke, graphitic shale, andesite, trondhjemite-granodiorite |
| Copper Lode MA (1) | Uchi | Uchi–Confederation | Confederation | bimodal | basalt, FIII-rhyolite, FII-rhyolite |
| Copper Lode E (1) | Uchi | Uchi–Confederation | Confederation | bimodal | basalt, FIII-rhyolite, FII-rhyolite |
| South Bay (1) | Uchi | Uchi–Confederation | Confederation | bimodal cauldron complex | basalt, andesite, FIII-rhyolite, FII-rhyolite |

Table 22.1. General characteristics of Archean volcanic–associated, massive, base metal sulphide deposits in the Superior Province of Ontario.

| Name (Reference)[1] | Age (Ma) | Footwall Rock | Cu wt.% | Zn wt.% | Pb wt.% | Ag oz/ton | Au oz/ton | Resource[2] Tonnes ($10^3$) |
|---|---|---|---|---|---|---|---|---|
| Stralak (1,9) | * | felsic tuffs and siliceous sedimentary rocks | 0.32 | 3.18 | * | 0.68 | * | 330 |
| Lake Geneva (1,9) | * | felsic tuffs and siliceous sedimentary rocks | * | 9.21 | 3.34 | 22.0 | 0.08 | 73 |
| Kam Kotia (1,3) | 2705 | mafic volcanic rocks | 1.09 | 1.03 | * | 0.11 | 0.001 | 6846 |
| Jameland (1,3) | 2705 | mafic volcanic rocks | 0.99 | 0.88 | * | 0.11 | 0.001 | 461.8 |
| Canadian Jamieson (1,3) | 2705 | mafic fragmental rocks | 2.3 | 3.5 | * | 0.88 | 0.009 | 826 |
| Genex (3) | 2705 | pillowed, massive, fragmental tholeiitic basalt | 21 to 27 | * | * | * | * | 0.242 |
| Potter (1) | 2715 | basalt fragmental rocks | 2.05 | 1.67 | * | * | * | 544 |
| Kidd Creek (1,2) | 2717 | rhyolite | 2.66 | 6.47 | 0.24 | 2.9 | * | 129600 |
| Winston Lake (4) | 2800 to 2700 | rhyolite | 1.00 | 15.6 | * | 0.99 | 0.03 | 3076 |
| Willroy (1,6,7) | 2800 to 2700 | mafic volcanic rocks? | 1.64 | 2.84 | * | 0.81 | * | 4355 |
| Big Nama Creek (1,6,7) | 2800 to 2700 | mafic volcanic rocks? | 0.83 | 4.16 | 0.02 | 1.04 | * | 181 |
| Willecho (1,6,7) | 2800 to 2700 | mafic volcanic rocks? | 0.5 | 4.43 | 0.18 | 1.98 | * | 2163 |
| Geco (1,6,7) | 2800 to 2700 | mafic volcanic rocks? | 1.86 | 3.45 | 0.15 | 1.46 | * | 50337 |
| Headway–Coulee area (1,8) | 2764 | felsic ash and tuff | 1.45 | 4.6 | * | 2.0 | * | 469 |
| Mattabi (1,5) | 2735 | intermediate to felsic pyroclastic rocks | 0.74 | 8.28 | 0.85 | 3.31 | 0.007 | 13805 |
| F–Group (5) | 2735 | felsic ash-flow tuff | 0.64 | 9.51 | 0.58 | 1.92 | * | 418 |
| Sturgeon Lake (1,5) | 2735 | fragmental rhyolite | 2.55 | 9.17 | 1.21 | 5.22 | 0.019 | 2510 |
| Lyon Lake and Creek Zone (1,5) | 2735 | fragmental rhyolite | 1.26 | 8.67 | 0.99 | 4.5 | 0.01 | 3487 |
| Copper Lode MA (1) | 2740 | rhyolite | 1.56 | * | * | 1.08 | * | 386 |
| Copper Lode E (1) | 2740 | rhyolite | 0.66 | 4.8 | * | 0.43 | * | 250 |
| South Bay (1) | 2740 | rhyolite-quartz-feldspar porphyry dome | 1.6 | 11 | * | 2.4 | * | 1570 |

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[1] *References: 1– J.M. Franklin, Geological Survey of Canada, unpublished data; 2– Brisbin et al. 1990; 3– Binney and Barrie 1990; 4– Severin et al. 1990; 5– Morton et al. 1990; 6– Williams et al. 1990, p. 9; 7– Friesen et al. 1982; 8– Osterberg et al. 1987; 9– Card and Innes 1981.*

[2] *Resource estimate may include past production and proven and probable reserves.*

* *not known*

**Table 22.2.** Summary of selected trace element characteristics of felsic metavolcanic rocks in the Superior Province (*after* Lesher et al. 1986; Ujike and Goodwin 1987; Barrie 1990).

| | **Rhyolite Classification** | | | |
|---|---|---|---|---|
| **Element or Ratio** | FI | FII | FIIIa | FIIIb |
| $(La/Yb)_N$ | 6 to 34 | 2 to 6 | 1 to 4 | 1 to 4 |
| Eu/Eu$^*$ | 0.87 to 2.0 | 0.35 to 1.4 | 0.37 to 0.94 | 0.20 to 0.61 |
| Zr/Y | 9 to 31 | 6 to 11 | 4 to 7 | 2 to 6 |
| Zr (ppm) | 90 to 275 | 95 to 430 | 170 to 370 | 190 to 730 |
| Y (ppm) | 3 to 30 | 7 to 55 | 25 to 70 | 70 to 240 |
| Sc (ppm) | 1.5 to 21 | 4.5 to 25 | 7 to 20 | 1.5 to 10 |
| Sr (ppm) | 100 to 620 | 45 to 265 | 5 to 100 | 7 to 210 |
| $TiO_2$ (wt%) | 0.15 to 0.65 | 0.15 to 0.90 | 0.20 to 1.0 | 0.10 to 0.75 |
| **Example Rock Successions** | | | | |
| Not Known to Contain VMS Deposits | Skead<br>Lake of the Woods<br>Shoal Lake<br>Kakagi Lake<br>Confederation Lake (Cycles 1, 2, upper 3)<br>Sturgeon Lake (Cycle 3) | Wabigoon Lake<br>Sturgeon Lake (Cycle 2)<br>Confederation Lake (Cycle 2) | | |
| Contain VMS Deposits | | Sturgeon Lake (Cycle 1a, 1b) | Noranda (Cycles 2, 3, 4) | Kamiskotia<br>Kidd Creek<br>Matagami<br>Noranda (Northwest rhyolite)<br>Confederation Lake (Cycle III) |

Thus, a component of synvolcanic faulting may be indicated (Coad 1976).

Despite the spatial association with the tholeiitic and komatiitic rock, the nickel tenor of the sulphide mineralization averaged only 0.03 weight % (Coad 1976, p.172). The massive ore was zoned from a zinc-rich top to a copper-rich base (Coad 1976), and the average copper:zinc ratio was close to, or less than, 1 (Coad 1976). Nickel:cobalt ratios of the ore were less than 1 (Coad 1976). The deposit was copper-rich and zinc-poor in proximity to the chlorite alteration zone, consistent with the premise that the chlorite alteration represented a feeder pipe (Coad 1976).

The presence of an inferred heat source (Centre Hill Complex), a conduit for hydrothermal fluids (fault at the contact between Centre Hill Complex and the komatiite paleo-lava lake), and a permeable aquifer (tholeiitic fragmental unit) in which the disseminated and massive sulphide minerals occur, are characteristic of a depositional environment common to many VMS deposits. However, the dominance of komatiite and tholeiite in the Potter mine setting, and the virtual absence of intermediate to felsic volcanic rocks, distinguishes this deposit setting from the many other VMS deposits in the Superior Province.

## DIVERSE KOMATIITE, THOLEIITE, ANDESITE AND RHYOLITE SUCCESSIONS

The giant, polymetallic Kidd Creek VMS deposit (Walker and Mannard 1974; Walker et al. 1975) occurs at the western end of the Kidd–Munro assemblage, Abitibi greenstone belt (Jackson and Fyon, this volume). This part of the Kidd–Munro assemblage consists of diverse rock types, including ultramafic, pyroxenitic, and basaltic komatiite, variolitic, massive, and pillowed tholeiitic basalt, magnesium-rich tholeiite, icelandite, thin units of 2716 ± 4 million-year-old (Barrie and Davis 1990) FIII-rhyolite, calc-alkalic metavolcanic rocks, and associated layered, tholeiitic and komatiitic intrusions (BVSP 1981, Chapter 1.2; Leahy and Ginn 1961a, 1961b; Arndt and Nesbitt 1982; Brisbin et al. 1990; Jackson and Fyon, this volume). No other significant VMS deposits in the Superior Province of Ontario occur within such a diverse komatiite-tholeiite-rhyolite assemblage.

The Kidd Creek deposit consists of several lenses, which occur at the top of a locally thickened, felsic metavolcanic succession, along the contact between subaqueous felsic and overlying mafic metavolcanic rocks (Coad 1985; Brisbin et al. 1990). Ultramafic rocks of volcanic or plutonic origin occur in the deposit footwall (Brisbin et al. 1990). A subvolcanic intrusion has not been positively identified in the Kidd Creek deposit area (Brisbin et al. 1990). The geometry of volcanic rocks and the sulphide deposit defines a large S-shape (Brisbin et al. 1990).

Coincident with the nose of the S-fold in the Kidd Creek deposit area is a thickened (greater than 300 m) accumulation of FIII-rhyolite (Brisbin et al. 1990). The country rocks and deposit are complexly folded and faulted (Brisbin et al. 1990). The footwall rhyolite and sulphide deposit are tabular and prolate in shape, elongate down-dip, and plunge steeply to the northeast, parallel to an intersection lineation between steeply dipping southeast- and east-striking cleavages (Brisbin et al. 1990). Footwall alteration and stringer mineralization are best developed along the vertical axis of the thickened felsic metavolcanic succession (Brisbin et al. 1990).

## SUBVOLCANIC INTRUSIONS

Subvolcanic intrusions, ranging in composition from gabbro to anorthositic gabbro or trondhjemite to granodiorite, lie up to several kilometres stratigraphically beneath many clusters of VMS deposits (Campbell et al. 1981; Franklin et al. 1981). Some of these intrusions are cogenetic and coeval with the associated metavolcanic rocks (Goldie 1976; Poulsen and Franklin 1981; Franklin 1986), and may have provided heat and metals to the hydrothermal system (Campbell et al. 1981).

## ALTERATION

Country rock in proximity to VMS deposits is altered. Typically, proximal to deposits, rocks are depleted in $Na_2O$, and enriched in MgO, $K_2O$ and FeO (Riverin and Hodgson 1980; Hall 1982; Knuckey et al. 1982; Knuckey and Watkins 1982; Gibson et al. 1983; Osterberg et al. 1987; Morton et al. 1990). These chemical changes are reflected in the resultant metamorphic mineralogy of the rock (Figure 22.4). Two styles of alteration are recognized: 1) regional semiconformable; and 2) a pipe-like form.

### Regional Semiconformable Alteration

At a camp scale, a regionally extensive, or lower semiconformable, alteration occurs distant from, and stratigraphically below, VMS deposits (*see* Figure 22.4; Morton and Franklin 1987). This alteration style can be developed up to 3 km below the paleo-sea floor on which the sulphides precipitated (MacGeehan and MacLean 1980; Gibson et al. 1983). It is areally extensive and can be hundreds of metres in thickness (Morton and Franklin 1987). An important factor contributing to the geometry and distribution of this alteration style is the permeability of the country rock (Gibson and Severin 1991); the lower semiconformable alteration style is better developed in fragmental rocks than in massive, effusive rocks (Gibson and Severin 1991). Lower semiconformable alteration zones associated with deposits in the Sturgeon Lake camp extend closer to the paleo-sea floor on which the sulphide mounds accumulated, compared to the lower semiconformable alteration associated with many deposits in the Noranda camp (Morton and Franklin 1987).

Mineral assemblages belonging to the lower semiconformable alteration type are spatially associated with deposits in the Sturgeon Lake camp, and consist of (*see* Figure 22.4): 1) sericite-quartz-calcite or dolomite; and 2) iron-rich chlorite, iron-bearing carbonate, chloritoid, and andalusite-kyanite (Groves 1984; Morton 1984; Morton and Franklin 1987). In the Noranda and Mattagami camps, lower semiconformable alteration comprises both silicified and spilitized (epidote and quartz with or without albite, actinolite, and magnesium chlorite) rock (MacGeehan and MacLean 1980; Gibson et al. 1983).

### Pipe Alteration

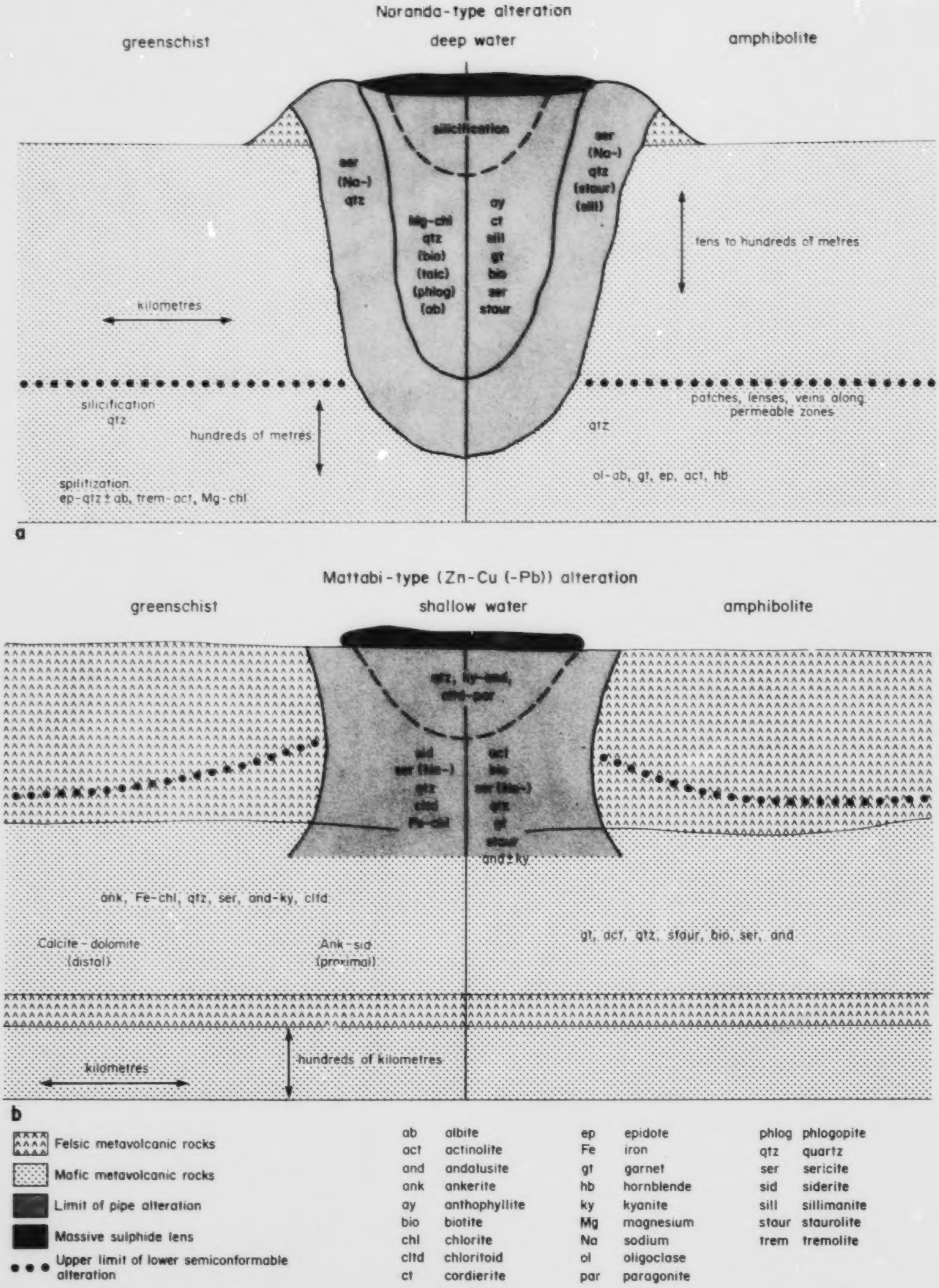
Alteration pipes are generally well developed in footwall rocks immediately beneath many deposits, and may cut across the regional semiconformable alteration (*see* Figure 22.4). Pipe alteration is well developed beneath many deposits in the Noranda and Mattagami camps (Riverin and Hodgson 1980; Franklin et al. 1981; Costa et al. 1983; Franklin 1986). Alteration pipes associated with the deposits in the Sturgeon Lake area (*see* Figure 22.4) are less pronounced than those associated with deposits in the Noranda camp (Morton and Franklin 1987). In proximity to deposits in the Sturgeon Lake camp, pipe alteration can extend for several hundred metres beneath the deposit and merge with the regionally extensive lower semiconformable alteration (Morton and Franklin 1987).

Alteration pipes, associated with greenschist-facies deposits in the Sturgeon Lake camp, consist of siderite, chloritoid, pods of sodium-rich sericite, iron chlorite (iron ripidolite or chamosite), andalusite and kyanite, localized in irregular pods or in fractures (*see* Figure 22.4; Morton and Franklin 1987). Silicified rock can occur immediately below the deposits. Alteration pipes associated with many greenschist-facies deposits in the Noranda camp are zoned from an inner magnesium-rich chlorite and quartz assemblage, to an outer sericite and quartz assemblage with, or without, chlorite (Riverin and Hodgson 1980; Franklin et al 1981; Franklin 1986). Talc is common beneath the Mattagami Lake deposit (Roberts and Reardon 1978; Costa et al. 1983). Where metamorphosed to amphibolite facies, the inner zones may consist of garnet, anthophyllite and cordierite, with varying amounts of iron chlorite, talc, phlogopite, and sillimanite or andalusite (Figures 22.5a and 22.5b) and the outer zone may consist of quartz, sodium-rich sericite, staurolite and smaller amounts of sillimanite (Morton and Franklin 1987).

## TEMPORAL CONSTRAINTS

All major VMS deposits in the Superior Province occur within volcanic assemblages that are younger than 2.8 billion years old (Figure 22.6). Within the 2.7 billion-year-old greenstone belts, VMS deposits occur within metavolcanic assemblages that range in age from 2738 to 2700 million years old, although older (2750 to 2775 million years old), unmineralized metavolcanic assemblages can also occur. Within a greenstone belt, not all volcanic assemblages of equivalent age contain VMS mineralization.

Detailed geochronological studies of alteration assemblages spatially associated with the Kidd Creek deposit indicate that potassium metasomatism and chlorite alteration occurred at 2624 ± 62 Ma; the deposit and host rocks are 2717 million years in age (Schandl and Davis 1989; Schandl et al. 1990). These younger alteration events are coeval with hydrothermal events spatially associated with "gold-only" deposits in the southern Abitibi


Noranda-type alteration
greenschist
deep water
amphibolite
silicification
ser
(Na-)
qtz
Mg-chl
qtz
(bio)
(talc)
(phlog)
(ab)
ay
ct
sill
gt
bio
ser
staur
ser
(Na-)
qtz
(staur)
(sill)
lens to hundreds of metres
kilometres
silicification
qtz
hundreds of metres
spilitization
ep-qtz ± ab, trem-act, Mg-chl
patches, lenses, veins along permeable zones
qtz
ol-ab, gt, ep, act, hb
a
Mattabi-type (Zn-Cu (-Pb)) alteration
greenschist
shallow water
amphibolite
qtz, ky-and, cltd-par
sid
ser (Na-)
qtz
chl
Fe-chl
act
bio
ser (Na-)
qtz
gt
staur
and ± ky
ank, Fe-chl, qtz, ser, and-ky, cltd
Calcite - dolomite
(distal)
Ank-sid
(proximal)
gt, act, qtz, staur, bio, ser, and
hundreds of kilometres
kilometres
b
Felsic metavolcanic rocks
Mafic metavolcanic rocks
Limit of pipe alteration
Massive sulphide lens
Upper limit of lower semiconformable alteration

| | | | | | | | |
|---|---|---|---|---|---|---|---|
| ab | albite | ep | epidote | phlog | phlogopite |
| act | actinolite | Fe | iron | qtz | quartz |
| and | andalusite | gt | garnet | ser | sericite |
| ank | ankerite | hb | hornblende | sid | siderite |
| ay | anthophyllite | ky | kyanite | sill | sillimanite |
| bio | biotite | Mg | magnesium | staur | staurolite |
| chl | chlorite | Na | sodium | trem | tremolite |
| cltd | chloritoid | ol | oligoclase | | |
| ct | cordierite | par | paragonite | | |

**Figure 22.4.** Synthesis of alteration assemblages and styles associated with volcanic-associated, massive, base metal sulphide (VMS) deposits in the Superior Province. Alteration developed under deeper water (Noranda-type) and shallower water (Mattabi-type) settings is indicated (*modified after* Morton and Franklin 1987).

**Figure 22.5.** Hydrothermally altered basaltic rock that was metamorphosed to amphibolite-grade facies. The alteration zone is spatially associated with the Winston Lake VMS deposit: **a)** bands and veinlets of garnet developed in a rock containing anthophyllite; **b)** large (up to 6 cm) crystals of anthophyllite, accompanied by garnet and sillimanite in an altered pillowed basalt.

greenstone belt, and medium- to high-grade metamorphism in the middle crust (*see* "GOLD-ONLY MINERALIZATION—Temporal Constraints, below). These data suggest that crustal-scale thermal events that occurred later than 2700 Ma are recorded in VMS and gold-only deposits. The effects of these younger crustal-scale thermal events were superimposed upon the earlier-formed VMS deposits and genetically related hydrothermal alteration assemblages. The significance of this late alteration event is not known.

## TECTONIC SETTING OF ARCHEAN VMS DEPOSITS

Possible tectonic settings in which VMS deposits formed in the Superior Province are constrained by the rock types of the host assemblages. Described below is the evidence for formation of VMS deposits in the Superior Province of Ontario in an extensional setting, analogous to some VMS deposits of Phanerozoic age (Cathles et al. 1983).

### Komatiite-Tholeiite Assemblages

A clue to the tectonic setting of the komatiite-tholeiite assemblages is provided by the presence of the icelandite and FIII-rhyolites (Campbell et al. 1982; Lesher et al. 1986; Barrie 1990). These distinctive rocks are part of a tholeiitic basalt–icelandite–FIII-rhyolite suite, derived by extreme plagioclase fractionation in high-level magma chambers (Macdonald et al. 1990). In modern settings, this suite is developed in rift environments, including: 1) Tertiary and Recent bimodal volcanic fields, such as in the Basin and Range Province, USA (Bacon and Metz 1984; Moyer and Esperanca 1989); and 2) off-axis settings along mid-oceanic ridges (e.g., eastern Iceland rift zone: Wood 1978; Sigurdsson and Sparks 1981; Macdonald et al., 1987, 1990). In Iceland, the basalt–icelandite–FIII-rhyolite suite is developed over "off-axis" fracture zones, where the basaltic crust is thicker, heat flow is lower, and basaltic magmas undergo extensive fractionation (Macdonald et al. 1990). In this setting, a heterogeneous rock assemblage, including several types of basalt, icelandite and rhyolite can occur (Macdonald et al. 1990).

By analogy, the presence of komatiites, tholeiites, chemically evolved basalts, icelandite and FIII-rhyolite in the Kidd–Munro assemblage may indicate evolution in an off-axis oceanic rift, back-arc or rifted arc setting.

The tholeiite-komatiite-dominated successions, such as that associated with the Potter Mine in the eastern part of the Kidd–Munro assemblage, differ from typical rock successions associated with many VMS deposits in the Superior Province. The predominance of mafic and ultramafic volcanic rocks, the abundance of tholeiitic effusive, fragmental and intrusive rocks associated with a komatiitic lava lake (Arndt 1986), the presence of a synvolcanic fault, the localization of chlorite alteration and disseminated sulphide mineralization along the fault zone, and the thickness and metal ratio variations in the massive sulphide zone collectively suggest that the Potter Mine VMS deposit formed in an oceanic setting by hydrothermal discharge along a synvolcanic fault (Coad 1976). This depositional setting may be analogous to an extensional, mid-oceanic ridge or primitive back-arc basin, dominated by tholeiite and komatiite igneous rocks (Jackson and Fyon, this volume).

### Bimodal Assemblages

Bimodal volcanic assemblages, such as the Blake River and Kamiskotia assemblages in the Abitibi greenstone belt (Jackson and Fyon, this volume) or the Confederation assemblage in the Uchi–Confederation greenstone belt (Stott, this volume), represent central volcanic complexes (Dimroth et al. 1985; Thurston et al. 1978) which developed as large shield volcanoes (Dimroth et al. 1985; Goodwin 1979; Thurston 1986; Gibson and Watkinson 1990). Many of the VMS deposits in the Superior Province formed in such a setting, following a period of cauldron collapse (Morton and Franklin 1991; Thurston 1986; Gibson and Watkinson 1990).

The presence of FIII-rhyolites in these assemblages is consistent with derivation of at least part of the volcanic succession by high-level fractionation of a tholeiitic magma chamber (Macdonald et al. 1990). Some assemblages, such as the Kamiskotia assemblage (Jackson and Fyon, this volume) are inferred to represent continental arcs (Barrie and Shirey 1991). Others, such as the Blake River assemblage, characterized by the presence of both calc-alkalic and tholeiitic basalt–(icelandite)–FIII-rhyolite suites, are inferred to represent rifted island arcs or back-arc basins (Ujike and Goodwin 1987; Gibson and Watkinson 1990; *see also* Williams, Stott, Thurston et al., this volume).

The VMS-associated cauldron-fill rhyolites in the Sturgeon Lake camp, designated as class FII (Condie 1981; Lesher et al. 1986), are chemically less evolved than the FIII-rhyolite present in the Noranda or Kidd Creek camps (Lesher et al. 1986). These chemically less evolved rhyolites may be transitional between those that resemble modern calc-alkalic rhyolites (FI: Condie 1981) and the FIII-rhyolites, and correspond to a tectonic setting transitional between island arc and rifted arc or back-arc basin (Lesher et al. 1986).

## ABUNDANCE OF VMS DEPOSITS IN THE ABITIBI GREENSTONE BELT

In the Superior Province, the Abitibi greenstone belt is noted, in part, for its abundance of VMS deposits (*see* Figure 22.3). The paucity of VMS deposits in 2.7 billion-year-old metavolcanic assemblages in other greenstone belts of the Wawa, Wabigoon and Uchi subprovinces may be related to: 1) the ensimatic arc setting of the southern Abitibi greenstone belt (Jackson and Sutcliffe 1990), which contrasts with the ensialic arc setting of some other greenstone belts, such as the Michipicoten (Sylvester et al. 1987); 2) poorly developed extensional tectonic settings in VMS-poor greenstone belts; and 3) the apparent dominance of assemblages that developed in a complex "micro-plate" tectonic setting (*see* Jackson and Fyon, this volume).

The Abitibi greenstone belt is characterized (Jackson and Fyon, this volume) by: 1) abundant komatiites and members of the tholeiite–icelandite–FIII-rhyolite suite; and 2) many small assemblages that may represent disparate coeval tectonic settings, such as arc, rifted arc, and back-arc basins. The diversity of areally restricted, coeval supracrustal assemblages, and the repetition of certain assemblage types in time and space (*see* Figure 11.36 *in* Jackson and Fyon, this volume) may indicate that the southern Abitibi greenstone belt is the product of a complex tectonic environment where "micro-plates" developed and interacted. An analogous modern tectonic setting may be represented by the Indonesian region, where many different "micro-plates" develop and interact. It is possible that the formation of VMS deposits was favoured in such an ensimatic tectonic setting.

Oceanic-like assemblages in greenstone belts of the eastern Wawa, Wabigoon and Uchi subprovinces do not contain abundant komatiitic flows or volcanic rocks of the basalt–icelandite–FIII-rhyolite suite (Lesher et al. 1986). These greenstone belts are dominated by tholeiitic basalt successions and FI-felsic metavolcanic rocks (Lesher et al. 1986; Blackburn et al., this volume). Certain aspects of the geological history of the Uchi Subprovince are analogous to the Phanerozoic evolution of the Andes Mountains (Stott and Corfu, this volume). Such a tectonic history differs in scale and complexity from the "micro-plate" tectonic development of the southwestern part of the Abitibi greenstone

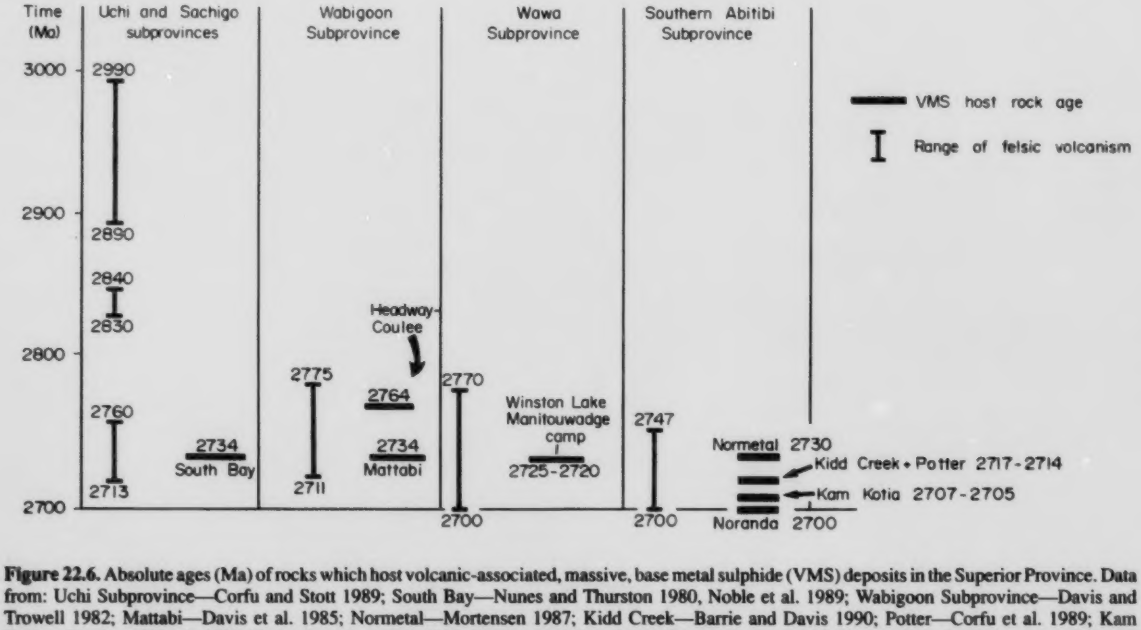


**Figure 22.6.** Absolute ages (Ma) of rocks which host volcanic-associated, massive, base metal sulphide (VMS) deposits in the Superior Province. Data from: Uchi Subprovince—Corfu and Stott 1989; South Bay—Nunes and Thurston 1980, Noble et al. 1989; Wabigoon Subprovince—Davis and Trowell 1982; Mattabi—Davis et al. 1985; Normetal—Mortensen 1987; Kidd Creek—Barrie and Davis 1990; Potter—Corfu et al. 1989; Kam Kotia—Barrie and Davis 1990; Noranda—Mortensen 1987, Davis, unpublished data; Winston Lake and Manitouwadge—Davis et al., unpublished data.

belt, and may account for the paucity of VMS deposits in the Uchi Subprovince.

The longevity of an hydrothermal system is dependent upon the rate of magma chamber refill, the cooling rate of the magma, the ease of external water influx and subsequent interaction with the magma, and the focussed or diffuse nature of hydrothermal fluid discharge (cf., Thurston 1986). The abundance of komatiite in the Abitibi greenstone belt may indicate melt generation in extensional oceanic rift settings overlying mantle regions that had higher temperatures compared to tholeiite-producing rifts. The result was the production of large volumes of hot, magnesian magma (McKenzie and Bickle 1988). The coexistence of komatiite with tholeiite, icelandite and FIII-rhyolite in the Abitibi greenstone belt indicates that high-level tholeiitic magma chambers formed and fractionated in the same general tectonic setting as that which existed when the komatiites were being erupted. In an oceanic regime, this was likely a rift setting. Heat flow in this setting, related to elevated mantle temperatures, may have been focussed along fewer axial and off-axial rifts. The initiation and maintenance of hot, metalliferous, hydrothermal systems was favoured in these tectonic settings where long-lived tholeiitic magma chambers are present high in the crust (Macdonald et al. 1990).

The longevity of an hydrothermal system is also a function of cauldron setting. Optimal conditions exist where resurgent volcanism is implicated because there was no catastrophic emptying of the magma chamber. This favoured the long-term release of latent heat during crystallization of the magma. Greenstone belts dominated by tholeiite and FI-dacite and rhyolite successions may have developed in oceanic settings underlain by lower-temperature mantle source regions where magmas of lower MgO content were produced (McKenzie and Bickle 1988). The FI-dacites and rhyolites appear to have developed in convergent, rather than extensional, settings (Sylvester et al. 1987; Barrie and Shirey 1991). Heat flow in the convergent tectonic setting in which the bimodal volcanic rocks formed may have been less focussed, which in turn, discouraged the development and maintenance of a hot hydrothermal system.

## Komatiitic-Associated Nickel-Copper-Platinum Group Element Sulphide Mineralization

Deposits of magmatic nickel-copper-platinum group element sulphide mineralization, spatially and genetically related to olivine-spinifex-bearing komatiites (Groves and Hudson 1981; Marston et al. 1981; Naldrett 1981; Ross and Travis 1981), occur within many Archean greenstone belts around the world, and account for about 25% of the world total nickel resource in deposits having greater than 0.8% nickel (Lesher 1989). These deposits are subdivided into komatiitic peridotite-hosted (Class I) and komatiitic dunite-hosted (Class II) classes, each, in turn, subdivided into strati-form (Type A) and stratabound (Type B) deposits (Lesher 1989). In these deposits, nickel is more abundant than copper, $[Ni/(Ni + Cu)] > 0.9$, and the mineralization occurs as massive, net-textured and disseminated pyrrhotite, chalcopyrite and pentlandite. The mineralization occurs near the base of ultramafic flows or within dunitic intrusions (Naldrett and Macdonald 1980). Where altered, the primary pyrrhotite-chalcopyrite-pentlandite assemblage can contain pyrite, millerite and heazlewoodite.

### DISTRIBUTION OF KOMATIITES AND ASSOCIATED NICKEL-COPPER-PLATINUM GROUP ELEMENT DEPOSITS

Within the Superior Province, komatiites are most common in the Sachigo Subprovince and the southern part of the Abitibi greenstone belt in the Abitibi Subprovince (Figure 22.7). Minor komatiite occurrences are known in the 2.8 billion-year-old fragment of the Michipicoten greenstone belt (Sage 1991); in the Wawa Subprovince west of Thunder Bay (Williams, Stott, Heather et al., this volume); in the southern and northeastern parts of the Wabigoon Subprovince (Thurston et al. 1986); and in the Red Lake greenstone belt in the Uchi Subprovince (Wallace et al. 1986). Komatiites are rare to absent in the western part of the Wabigoon Subprovince (Blackburn et al., this volume).

Komatiite-associated nickel-copper-platinum group element deposits occur principally in the southern part of the Abitibi greenstone belt, and as clusters of deposits in the Timmins, Ontario (Table 22.3) and Malartic, Quebec areas (*see* Figure 22.7; Franklin and Thorpe 1982). The Shebandowan deposit, in the western part of the Wawa Subprovince, occurs in both a deformed peridotitic intrusion and adjacent mafic metavolcanic rocks (Morton 1982). This may be a deformed, komatiite-associated nickel-copper-platinum group element deposit which has been locally remobilized into adjacent metavolcanic rocks. Similarly, the Thierry deposit in the Pickle Lake area may also represent a redistributed komatiitic nickel-copper deposit (Patterson and Watkinson 1984), although the deposit occurs in a deformed mafic-ultramafic sill (Sage and Breaks 1982).

Deposit clusters of Class IA deposits in the Timmins area include (Figure 22.8): 1) Frederick House Lake area (Frederick House Lake, Alexo and Dundonald deposits) in the Kidd–Munro assemblage; 2) southern Shaw Dome area (Langmuir, McWatters, Redstone and Hart deposits) in the Bowman or Eldorado assemblage; 3) Bartlett Township (Texmont deposit) in the Bowman assemblage; and 4) Sothman township (Sothman deposit), possibly in the Bowman or Halliday assemblages. All of the deposits in the Timmins area occur in, or spatially associated with, olivine spinifex-bearing komatiite flows and cogenetic sills (Coad 1977; Muir and Comba 1979; Green and MacEachern 1990). The deposits in the Frederick House Lake area occur within a common komatiitic stratigraphic unit (Green and MacEachern 1990). Deposits in the southern Shaw Dome and Bartlett and Sothman townships area also occur along the base of a single komatiitic unit (Green and MacEachern 1990) that may be stratigraphically correlative (Green and MacEachern 1990; Jackson and Fyon, this volume).

## CHARACTERISTICS OF THE DEPOSITIONAL ENVIRONMENT

The following is a partial list of features characteristic of the depositional environment of the deposits in the Timmins area (Green 1978, 1980; Green and Naldrett 1981; Lesher 1989; Green and MacEachern 1990).

1. The deposits occur in clusters in a restricted area, generally near a single basal contact of the most magnesian, lowermost flow of a komatiite sequence.
2. Mineralized komatiite flows generally have a thick (10 to 20 m) lower cumulate zone and aphyric and spinifex-textured upper zones; stratigraphically equivalent, unmineralized flows are generally thinner and less magnesian than the mineralized komatiite flows.
3. Except where remobilization can be inferred, higher-grade, nickel sulphide zones occur beneath, or at the base of, komatiitic flows that are very high in MgO (greater than 40 weight %).
4. Locally, the nickel sulphide mineralization grades laterally into iron-sulphide-bearing, interflow metasedimentary rocks, where nickel tenor drops.
5. Nickel-sulphide mineralization may also occur within mafic metavolcanic rocks adjacent to komatiites, such as at the Dundonald, Hart, Redstone and Langmuir No. 2 deposits; however, the mineralization can be traced laterally, directly back into the main mineralized horizon.
6. Replacement of the komatiite flows by serpentine or talc-carbonate is quite variable, and apparently unrelated to the major ore zones; however, the nickel tenor of the mineralization is increased as a consequence of alteration.
7. Economic concentrations of nickel sulphide mineralization are not confined to zones of penetrative deformation.

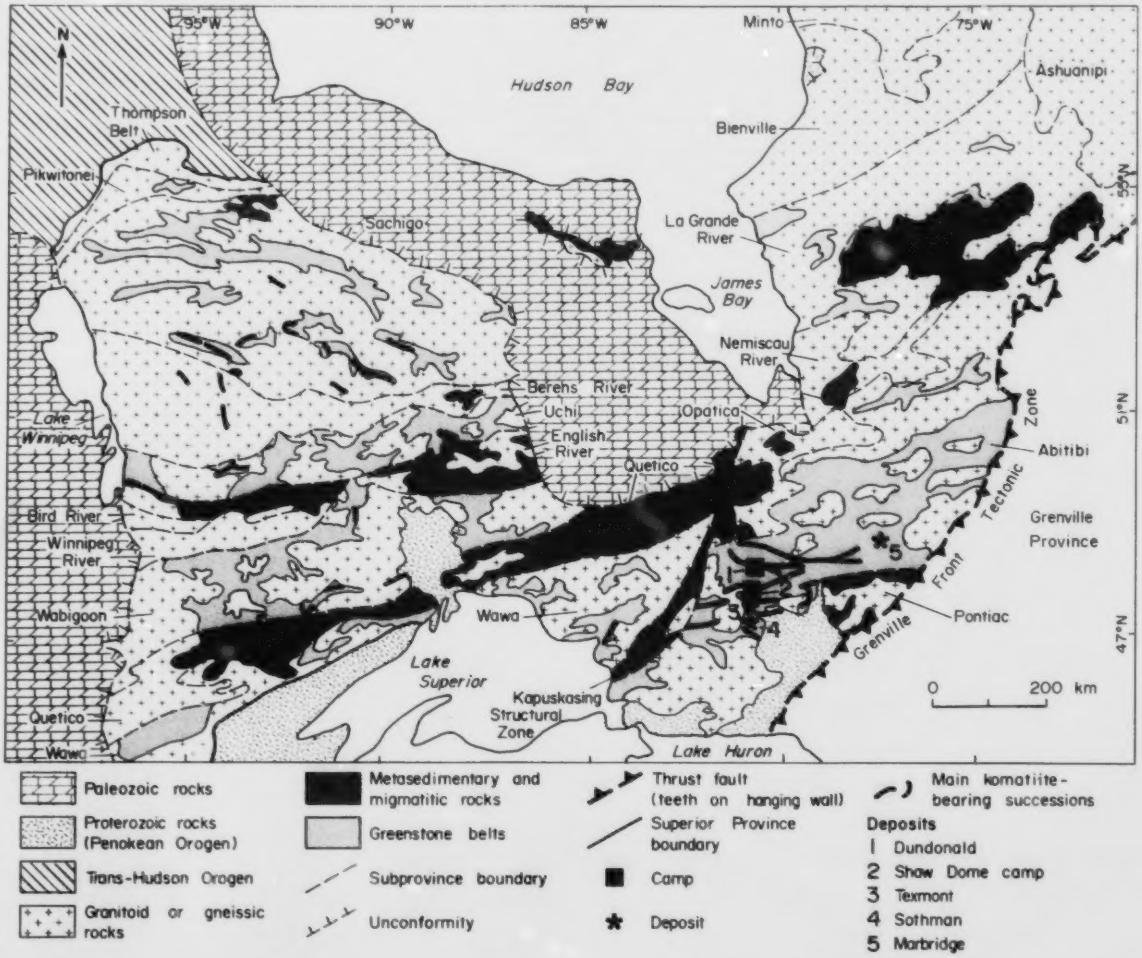

N
95°W
90°W
85°W
Minto
75°W
Ashuanipi
Hudson Bay
Thompson Belt
Pikwitonei
Bienville
55°N
Sachigo
La Grande River
James Bay
Nemiscau River
Berens River
Uchi
Opatica
Lake Winnipeg
English River
Quetico
Abitibi
51°N
Bird River
Winnipeg River
Grenville Province
Grenville Front Tectonic Zone
5
Wabigoon
Wawa
Pontiac
4
47°N
Lake Superior
Quetico
Wawa
Kapuskasing Structural Zone
Lake Huron
0
200 km
Paleozoic rocks
Proterozoic rocks (Penokean Orogen)
Trans-Hudson Orogen
Granitoid or gneissic rocks
Metasedimentary and migmatitic rocks
Greenstone belts
Subprovince boundary
Unconformity
Thrust fault (teeth on hanging wall)
Superior Province boundary
Camp
Deposit
Main komatiite-bearing successions
Deposits
1 Dundonald
2 Shaw Dome camp
3 Texmont
4 Sothman
5 Marbridge

**Figure 22.7.** Distribution of some komatiite-bearing successions and associated nickel-copper-platinum group element mineralization in the Superior Province. The simplified distribution of komatiite-bearing assemblages is illustrated for principal assemblages in Ontario.

**Table 22.3.** Komatiite-associated Ni-Cu-PGE deposits in the Timmins area, Abitibi greenstone belt (*Modified after* Luhta et al. 1990).

| Deposit (Reference[1]/ Note[2]) | Production Date | Resource[7,8] (t) | Ni% |
|---|---|---|---|
| **Frederick House Lake area (Kidd–Munro assemblage)** | | | |
| Alexo[3] | 1912–19<br>1933–44 | 51 530 | 3.93 |
| Dundonald[5] | no production | small tonnage | high grade |
| **Shaw Dome area (Bowman assemblage)** | | | |
| Langmuir No. 1[4] | 1973–77,<br>1990–91 | 110 000 | 1.7 |
| Langmuir No. 2[4] | 1973–77 | 997 900 | 1.5 |
| McWatters[4] | no production | 181 500<br>525 700 | 1.92<br>0.73 |
| Redstone[6] | 1989 | 257 000 | 2.4 |
| Hart[4] | no production | 770 000 | 0.09 |
| **other areas** | | | |
| Sothman | no production | 231 000<br>440 000 | 1.3<br>0.9 |
| Texmont | 1971–72 | 3 190 000 | 0.92 |

[1] *References: Naldrett 1964; Pyke 1978; Muir and Comba 1979; Green and Naldrett 1981; Robinson and Hutchinson 1982; Barnes & Naldrett 1987; Green and MacEachern 1990.*
[2] *Notes:* [3] *Noranda;* [4] *Timmins Nickel;* [5] *Falconbridge;* [6] *Timmins Nickel (51%)–BHP–Utah (49%);* [7] *Luhta et al. 1990.*
[8] *Resource is defined as past production (if any) plus remaining proven reserves.*

8. Sulphur isotope and selenium:sulphur ratios of sulphide minerals show that the sulphur within the deposits represents a mixture of magmatic and assimilated sulphur from sedimentary rocks.
9. Platinum group element concentrations are relatively uniform within deposits, but differ between clusters of deposits.

## TEMPORAL CONSTRAINTS

The age of the komatiite-associated nickel-copper-platinum group element mineralization is approximated by the age of spatially associated felsic metavolcanic rocks. In the Timmins area, these deposits occur in assemblages dating from 2727 to 2714 Ma. The Texmont and Sothman nickel sulphide deposits at the base of the Bowman assemblage, overlie the 2727 ± 1.5 million-year-old (Corfu et al. 1989) Bartlett assemblage (Jackson and Fyon, this volume). The age of deposits in the Halliday assemblage is not known. The similarity of rock types in this assemblage with the Skead, Blake River and Gauthier assemblages suggests that the Halliday assemblage could be 2700 million years old, although its spatial association with the Bartlett assemblage suggests it could be ~2730 million years old (Jackson and Fyon, this volume). The Langmuir, McWatters, Redstone and Hart deposits, along the south flank of the Shaw Dome in the Eldorado and Bowman assemblages, are inferred to have formed between 2727 and 2703 Ma (Nunes and Pyke 1980; Corfu et al. 1989; *see* discussion in Jackson and Fyon, this volume). Deposits in the Frederick House Lake area occur within the 2717 to 2714 million-year-old (Corfu et al. 1989; Barrie and Davis 1990) Kidd–Munro assemblage. Although the Class IA deposits in the Timmins area have been assigned to the lowermost, ultramafic unit of a regionally extensive upper metavolcanic supergroup (Green and Naldrett 1981; Jensen and Pyke 1982), the temporal constraints suggest that 3 ages of mineralization, 2727 Ma, 2715 Ma and 2700 Ma, may exist.

## KOMATIITE-THOLEIITE ASSEMBLAGES

In Western Australia, komatiites occur in both platform-phase and rift-phase greenstone belts (Groves and Batt 1984), and most class IA and IIB deposits occur in the rift-phase greenstone belts; some also occur in the platform-phase greenstone belts (Lesher 1989). This "rift" and "platform" terminology predates the tectonic affiliations suggested in this chapter for komatiite-bearing assemblages. In the Superior Province of Ontario, komatiites occur in at least 2 assemblage types:

1. platform-like assemblages (Thurston and Chivers 1990), found mainly in the Sachigo Subprovince
2. komatiite-tholeiite-dominant assemblages comprising diverse rock types, including komatiite, tholeiite, andesite, and rhyolite successions, such as the Kidd–Munro in the Abitibi greenstone belt (Jackson and Fyon, this volume)

Some of the komatiite-tholeiite-dominated assemblages have been defined as mafic plain assemblages by Thurston and Chivers (1990). More detailed descriptions of the deposits and their associated assemblages are provided to illustrate the various depositional settings.

### Platform-like Assemblages

The rock associations of platform assemblages were described and defined in Western Australia (Groves and Batt 1984), and the assemblage characteristics were subsequently applied to the Superior Province of Ontario (Wood, Thurston et al. 1986; deKemp 1987; Thurston et al. 1987; Thurston and Chivers 1990). Based primarily on reconnaissance study, combined with very localized detailed studies, it was found that platform assemblages in the Sachigo Subprovince consist of quartz arenite, sedimentary stromatolitic carbonate, iron formation and komatiitic or tholeiitic volcanic rocks (Wood, Thurston et al. 1986; deKemp 1987; Thurston et al. 1987). Platform assemblages appear to occur in lower parts of older, 3.02 to 2.95 billion-year-old greenstone belts and may be overlain by younger 2.75 to 2.7 billion-year-old greenstone belts (Wood, Thurston et al. 1986; Thurston 1986). Platform-like assemblages may also occur in the central Wabigoon region (Wabigoon Subprovince) (Thurston et al. 1987; Wilks and Nisbet 1985; Wood, Thurston et al. 1986). Komatiites associated with these sequences in the Superior Province appear to be devoid of komatiite-associated nickel-copper-platinum group element mineralization (Thurston and Chivers 1990).

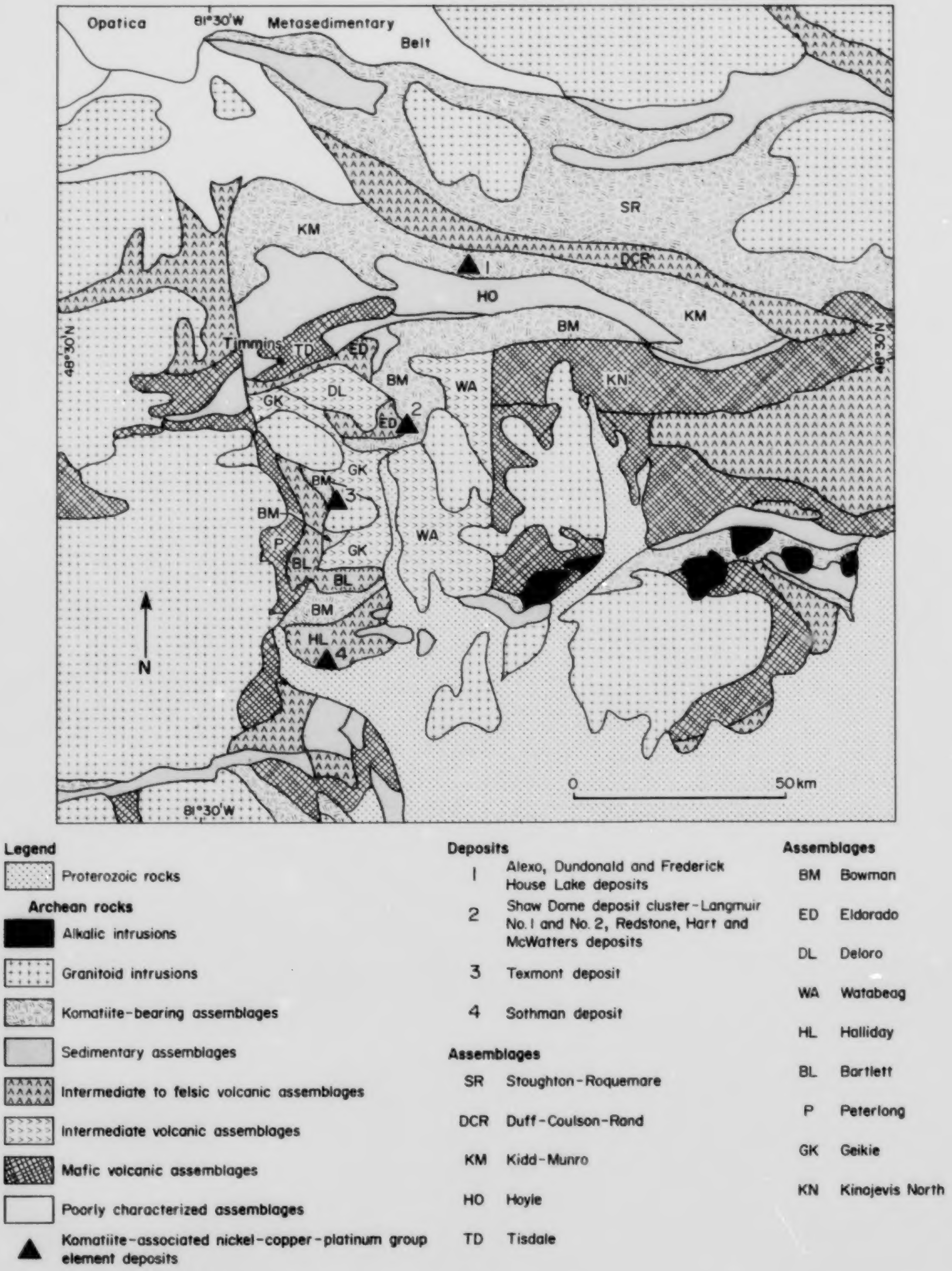

Opatica
81°30'W
Metasedimentary
Belt
SR
KM
1
DCR
HO
KM
BM
KN
48°30'N
48°30'N
Timmins
TD
ED
DL
BM
WA
GK
ED
2
GK
BM
3
BM
P
GK
WA
BL
BL
BM
HL
4
N
0
50km
81°30'W
Legend
Proterozoic rocks
Archean rocks
Alkalic intrusions
Granitoid intrusions
Komatiite-bearing assemblages
Sedimentary assemblages
Intermediate to felsic volcanic assemblages
Intermediate volcanic assemblages
Mafic volcanic assemblages
Poorly characterized assemblages
Komatiite-associated nickel-copper-platinum group element deposits
Deposits
1 Alexo, Dundonald and Frederick House Lake deposits
2 Shaw Dome deposit cluster-Langmuir No. 1 and No. 2, Redstone, Hart and McWatters deposits
3 Texmont deposit
4 Sothman deposit
Assemblages
SR Stoughton-Roquemare
DCR Duff-Coulson-Rand
KM Kidd-Munro
HO Hoyle
TD Tisdale
Assemblages
BM Bowman
ED Eldorado
DL Deloro
WA Watabeag
HL Halliday
BL Bartlett
P Peterlong
GK Geikie
KN Kinojevis North

**Figure 22.8a.** Distribution of komatiite-associated nickel-copper-platinum group element deposits in the Timmins area, southern Abitibi greenstone belt, illustrating the associated volcanic and sedimentary assemblages. For a complete listing of assemblage names and descriptions, *see* Figure 11.7 *in* Jackson and Fyon, this volume.

## Komatiite-Tholeiite Assemblages

A variety of rock associations characterizes these assemblages in the Abitibi greenstone belt, and several subdivisions of this assemblage type are proposed and described elsewhere (Jackson and Fyon, this volume). Two assemblages from the Abitibi greenstone belt that contain komatiite-associated nickel-copper-platinum group element mineralization, Kidd–Munro and Bowman (Jackson and Fyon, this volume), are described to illustrate 2 subtypes of this assemblage type.

### DIVERSE KOMATIITE, THOLEIITE, ANDESITE AND RHYOLITE ASSEMBLAGES

The Alexo, Dundonald and Frederick House Lake deposits, in the Kidd–Munro assemblage (*see* Figure 22.8), are associated with a succession of tholeiitic basalt, coarse rhyolitic fragmental rocks, cherty felsic lapilli tuff, spinifex-textured ultramafic and pyroxenitic komatiitic flows, magnesium-rich tholeiitic basalts and pyrite-bearing carbonaceous units (Naldrett 1964; Green and MacEachern 1990). The succession is intruded by the tholeiitic Dundonald Sill (Naldrett 1964; Naldrett and Mason 1968; Muir and Comba 1979).

Three habits of sulphide mineralization occur at the Dundonald deposit (Green and MacEachern 1990): 1) pyrrhotite and pyrite stringers or nodules and stratabound disseminated pyrrhotite in carbonaceous units; 2) disseminated, 2 to 5% pyrrhotite in the cumulate base of komatiitic flows; and 3) sulphide stringers or disseminated nodules in the dark green serpentinized peridotite. Pods and zones of massive and disseminated sulphide minerals occur at, or very near, the base of the thicker peridotite units. The stringer, nodule and disseminated sulphide mineralization, associated with graphite in fractures, occurs in ultramafic hyaloclastite flow-top units (Green and MacEachern 1990). Pyrrhotite, pentlandite, pyrite and minor amounts of millerite, godlevskite, heazlewoodite,

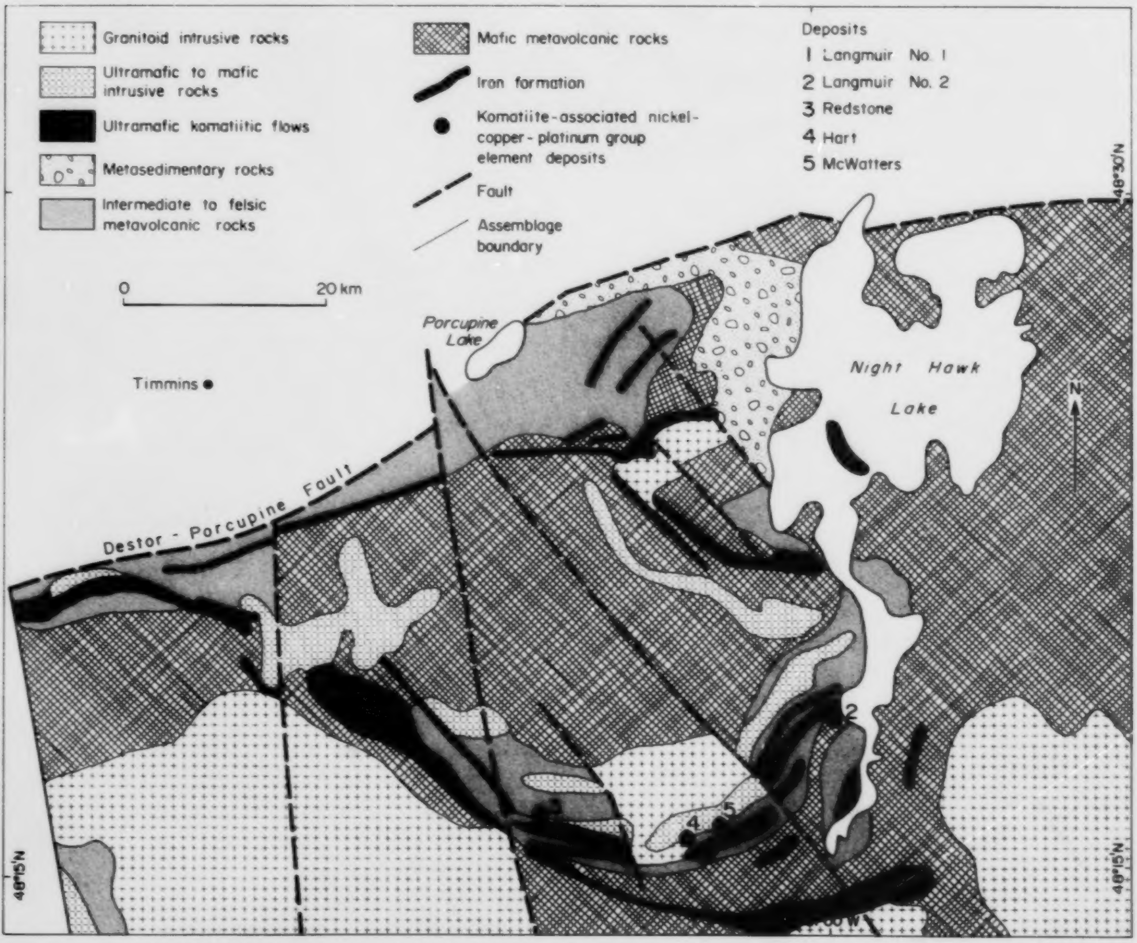

Granitoid intrusive rocks
Ultramafic to mafic intrusive rocks
Ultramafic komatiitic flows
Metasedimentary rocks
Intermediate to felsic metavolcanic rocks
Mafic metavolcanic rocks
Iron formation
Komatiite-associated nickel-copper-platinum group element deposits
Fault
Assemblage boundary
Deposits
1 Langmuir No. 1
2 Langmuir No. 2
3 Redstone
4 Hart
5 McWatters
0
20 km
48°30'N
48°15'N
Porcupine Lake
Timmins
Night Hawk Lake
N
Destor - Porcupine Fault
2
3
4
5

**Figure 22.8b.** Distribution of komatiite-associated nickel-copper-platinum group element deposits in the Timmins area, southern Abitibi greenstone belt, illustrating the associated volcanic and sedimentary assemblages.

gersdorffite, bravoite, sphalerite and chalcopyrite are the common sulphide minerals (Green and MacEachern 1990).

The Alexo deposit (*see* Figure 22.8) occurs in a steeply dipping footwall embayment, associated with a komatiitic flow having a dunitic core and pyroxene-rich peridotite margins (Green and MacEachern 1990). This ultramafic flow is correlative with pyroxene-rich flows which occur in pillowed metavolcanic rocks to the west, stratigraphically above the Dundonald Sill. The footwall rocks are pillowed basalts, interlayered with hyaloclastites and spinifex-bearing komatiitic flows (Green and MacEachern 1990). Massive sulphide lenses, in sharp contact with overlying, net-textured sulphides, consist of pyrrhotite, pentlandite, magnetite and chalcopyrite, and occur along the basal komatiite contact (Green and MacEachern 1990). Massive sulphides penetrate up to 5 m into the metavolcanic footwall (Green and MacEachern 1990).

### KOMATIITE-THOLEIITE-DOMINATED SUCCESSIONS

Some of these successions have been defined as mafic plains (Thurston and Chivers 1990). In the Abitibi greenstone belt, assemblages, such as the Stoughton–Roquemaure (Jensen and Langford 1985) and Bowman (Goodwin 1979), consist of komatiites and iron- and magnesium-tholeiitic flows with a rare, minor felsic metavolcanic component. South of Timmins (*see* Figure 22.8), the Bowman assemblage contains several komatiite-associated nickel-copper-platinum group element deposits (*see* Table 22.3; Pyke 1970; Coad 1977; Goodwin 1979; Robinson and Hutchinson 1982; Green and MacEachern 1990; Jackson and Fyon, this volume): 1) Redstone, Hart (Tontine), McWatters, Langmuir No. 1 and Langmuir No. 2 deposits in the Eldorado–Langmuir area along the south margin of the Shaw Dome; 2) Texmont, south of the Shaw Dome; and 3) Sothman, possibly at the base of the Bowman assemblage, on the south side of the Halliday assemblage.

The Bowman assemblage consists principally of komatiite and tholeiite. Massive, thick (25 m) spinifex-textured komatiites, rich in MgO (greater than 40 weight %), occur at the base of the Bowman assemblage; the nickel-copper-platinum group element sulphide deposits occur at the base of this flow unit (Green and MacEachern 1990). Locally, the basal komatiite unit of the Bowman assemblage is interlayered with the felsic to intermediate metavolcanic rocks and sulphidic interflow metasedimentary units of the underlying Eldorado or Bartlett assemblages (Pyke 1970, 1982; Green 1978; Green and MacEachern 1990; Jackson and Fyon, this volume).

Massive and disseminated mineralization consists of pyrrhotite and pentlandite with minor millerite (Green and MacEachern 1990). Pods of very high nickel tenor (greater than 20 weight %) are associated with zones where the adjacent serpentinized komatiite has been replaced by talc and carbonate (Green and MacEachern 1990). The nickel sulphide mineralization locally displays a lateral gradation into sulphide-bearing metasedimentary rocks (Green and MacEachern 1990). The apparent localization of the deposits along linear topographic depressions suggests a paleotopographic control on the siting of the nickel-copper-sulphur deposits (Green and Naldrett 1981; Green and MacEachern 1990).

## DEPOSITIONAL MODEL

Lava channels, magma chemistry and sulphide-bearing metasedimentary rocks are essential components of the depositional model (Figure 22.9) for komatiite-associated nickel-copper-platinum group element deposits (Ross and Hopkins 1975; Lesher et al. 1982, 1984; Lesher 1989). Voluminous eruption of komatiites into a rift setting formed laterally continuous lava channels. Thermal erosion beneath the komatiite lava channel resulted in assimilation of metasedimentary or felsic metavolcanic footwall rocks (Arndt and Jenner 1986; McNaughton et al. 1988). Flows that assimilated large amounts of sulphide-rich footwall rocks achieved sulphur saturation. A sulphide liquid segregated to form the classical sequence of massive, matrix and disseminated sulphide minerals at the base of the lowermost flow of the ultramafic succession (Lesher 1989). Flows characterized by smaller amounts of assimilation formed only disseminated nickel sulphides (Lesher 1989). Within a lava channel, komatiite-associated nickel-copper-platinum group element sulphide deposits were generally localized in, or over, footwall embayments (Lesher 1989). The embayments reflect a combination of structural deformation (Cowden 1988), thermal erosion (Evans et al. 1988; Frost and Groves 1988), and volcanic topography (Green and Naldrett 1981). Liquid sulphide accumulated in embayments, although some flanking mineralization also formed. Remobilization of sulphide minerals into the adjacent wall rock may have occurred by redistribution of low-viscosity sulphide liquid, or during post-solidus structural deformation.

In the Timmins area, sulphur isotopic values of sulphide minerals from the nickel-copper-platinum group element ores range widely and include non-magmatic values, consistent with the incorporation of sediment-derived sulphur into the magmatic system (Green and Naldrett 1981; Green and MacEachern 1990). Chromite in mineralized komatiite contains greater than 0.6 atomic % Zn, attributed to assimilation of zinc-rich metasedimentary rock (Lesher 1989). Chromite from unmineralized komatiite contains less than 0.6 atomic % Zn (Lesher 1989).

## TECTONIC SETTING OF KOMATIITE-ASSOCIATED NICKEL-COPPER-PLATINUM GROUP ELEMENT MINERALIZATION

Direct Phanerozoic analogues of komatiites, such as Gorgona Island in the Pacific Ocean (Echeverria 1982), are rare; hence, rock association and geochemical characteristics of associated rocks in Archean assemblages must be used to infer the tectonic setting of Archean komatiites.

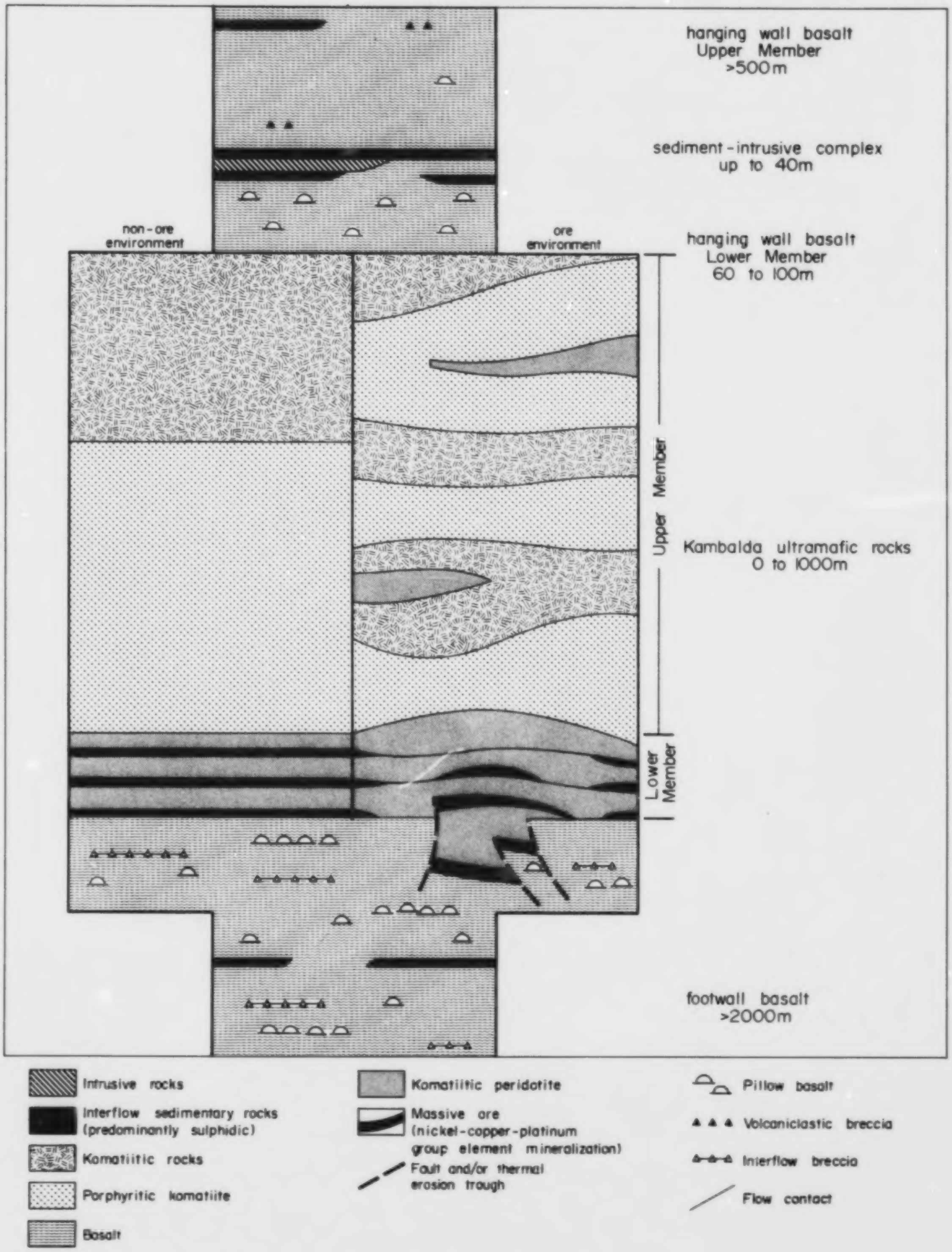

hanging wall basalt
Upper Member
>500m
sediment-intrusive complex
up to 40m
non-ore environment
ore environment
hanging wall basalt
Lower Member
60 to 100m
Upper Member
Kambalda ultramafic rocks
0 to 1000m
Lower Member
footwall basalt
>2000m
Intrusive rocks
Interflow sedimentary rocks (predominantly sulphidic)
Komatiitic rocks
Porphyritic komatiite
Basalt
Komatiitic peridotite
Massive ore (nickel-copper-platinum group element mineralization)
Fault and/or thermal erosion trough
Pillow basalt
Volcaniclastic breccia
Interflow breccia
Flow contact

**Figure 22.9.** Depositional model for komatiite-associated, nickel-copper-platinum group element mineralization (*modified after* Gresham and Loftus-Hills 1981).

## Platform-like Assemblages

The association of komatiites and tholeiites at the top of platform-like assemblages (Groves and Batt 1984; Thurston and Chivers 1990) may signify initiation of intracratonic rifting (Thurston, this volume). In the Superior Province, komatiite-associated nickel-copper-platinum group element deposits have not yet been discovered in this setting (Thurston and Chivers 1990).

## Komatiite-Tholeiite-Dominated Assemblages

Komatiites occur in komatiite-tholeiite assemblages, such as the Bowman, Tisdale and Stoughton–Roquemaure assemblages in the Abitibi greenstone belt. They are defined as mafic plain-like assemblages (Thurston and Chivers 1990). These komatiites may have been erupted onto rifted oceanic crust (Groves and Batt 1984; Thurston and Chivers 1990).

In the southwestern Abitibi greenstone belt, komatiite-associated nickel-copper-platinum group element deposits occur in only the Bowman assemblage. A clue to the tectonic setting represented by the Bowman assemblage is provided by its stratigraphic contact with the underlying, fractionated rhyolites of the Eldorado and Bartlett assemblages. These rhyolites chemically resemble modern calc-alkalic rhyolites (Goodwin 1979; Pyke 1982), and, hence, may have evolved in an island arc setting. The presence of komatiites stratigraphically overlying the felsic metavolcanic rocks indicates that the setting may be analogous to: 1) island arc, or 2) rifted island arc.

Archean komatiites that formed in an oceanic or island arc setting have been distinguished on the basis of their REE and Sm-Nd isotope systematics (Krogstad et al. 1989; Rajamani et al. 1985). Komatiites that are depleted in light rare earth elements (LREEs) evolved in an ocean ridge setting (Rajamani et al. 1985). Light rare earth element-enriched komatiites with high, positive epsilon neodymium values were derived from a mantle depleted in LREEs that was enriched in LREEs by subduction-driven mantle metasomatism just before formation of komatiites (Rajamani et al. 1985). This komatiite-type was erupted into a rifted ocean island or island arc environment (Rajamani et al. 1985). The lack of comprehensive chemical data prohibits a more rigorous classification of the tectonic setting in which the komatiites of the Bowman assemblage erupted.

## Diverse Komatiite-Tholeiite Assemblages

In the diverse komatiite-tholeiite assemblages, such as the Kidd–Munro (Jackson and Fyon, this volume), the presence of icelandite and high-silica rhyolite with komatiite is consistent with komatiite eruption in an extensional back-arc setting (Crecraft et al. 1981; Bacon and Metz 1984; Moyer and Esperanca 1989) or hot spot, mid-ocean rift setting (Sigurdsson and Sparks 1981; Macdonald et al. 1987, 1990; Wood 1978). The komatiite-associated nickel-copper-platinum group element deposits of the Dundonald area appear to have formed in such an extensional oceanic setting.

## Summary

Although there are indications that petrogenetically different types of komatiites occur in some Archean greenstone belts (Krogstad et al. 1989; Rajamani et al. 1985), komatiite-associated, nickel-copper-platinum group element deposits in the Abitibi greenstone belt appear to have formed in rifted island arc, back-arc, or mid-ocean rift settings. We recognize that further chemical characterization of the komatiites is required to constrain possible eruptive environments.

# Iron Formation

Two major types of iron-rich sedimentary rocks are recognized (Gross 1965; James 1966; Kimberley 1978): 1) ironstones, which are noncherty, contain greater than 15% iron, and are largely confined to the Phanerozoic; and 2) iron formation, which is typically laminated with alternating iron-rich and iron-poor layers, and is largely, but not exclusively, Precambrian in age. The mineralogy of iron-rich phases determines the facies classification (James 1954): 1) oxide facies—magnetite and/or hematite; 2) carbonate facies—siderite and/or ankerite; 3) sulphide facies—pyrite and/or pyrrhotite; and 4) silicate facies. The iron silicates comprise primary greenalite, minnesotaite and possibly stilpnomelane; metamorphic minerals include chamosite and other iron chlorites and iron amphiboles. The iron-poor layers in all facies of iron formation include chert, wacke, argillite and carbonate metasedimentary rocks.

In the Superior Province of Ontario, major iron formation units occur in most subprovinces (Figure 22.10).

## PALEOENVIRONMENTAL CLASSIFICATION

The Algoma- and Superior-types of iron formation described by Gross (1965) have been supplanted by the paleoenvironmental classification of Kimberley (1978):

1. shallow-volcanic-platform iron formation deposited in a shallow-water setting above volcanic units on a platform or in a lagoon close to wave base
2. metazoan-poor, areally extensive, chemical-sediment-rich, shallow-sea iron formation (iron formation described by James 1966) forming extensive Paleoproterozoic units deposited in a shallow continental shelf with little relief adjacent to a continent, but with trapping of terrigenous sediments outside the basin—the Hamersley iron formation is of this type, but is of Archean age
3. sandy, clayey and oolitic, shallow inland-sea iron formation (ironstone described by James 1966) deposited in highly agitated shallow water
4. deep-water iron formation, deposited in a deep-water volcanic environment and characteristically iron-poor, associated with ferriferous chert; units are generally quite thin and commonly form the uppermost unit of turbiditic wacke units

The tectonic settings of Kimberley (1978) have been amalgamated with additional, recently described depositional environments of Archean supracrustal sequences (*after* Thurston and Chivers 1990), in Table 22.4. The presence of iron formation is a sensitive indicator of paleoenvironment; hence it can be used to constrain the tectonic settings of associated rocks in some assemblages of the Superior Province.

## Arc Environment

The following are major characteristics of iron formation in arc settings (*see* Table 22.4):

1. Arc-related iron formation is separable into a deep- and a shallow-water environment.
2. The shallow-water type, where undeformed, contains oolitic primary structures (Lambert 1976), similar to many undeformed Proterozoic ironstones.
3. Alteration is common in the underlying volcanic rocks (e.g., magnesium and $CO_2$ metasomatism at Wawa, Morton and Nebel 1984).
4. This shallow-water, arc-related type may be the lateral equivalent of volcanogenic massive sulphide deposits. An example is the thin oxide-facies cherty metasedimentary rock known as the "key tuffite" in the Noranda area (Spence and de Rosen-Spence 1975). Ontario examples are the lateral relationship between iron formation and an orebody at Sturgeon Lake (Morton et al. 1990), and a similar relationship between iron formation and the Geco VMS deposit at Manitouwadge (Pye 1960).
5. The deeper-water iron formation in the arc environment may be separable into a proximal sulphide facies with graphitic shale background sediment, whereas the more distal units are oxide facies with turbidites (Shegelski 1978).

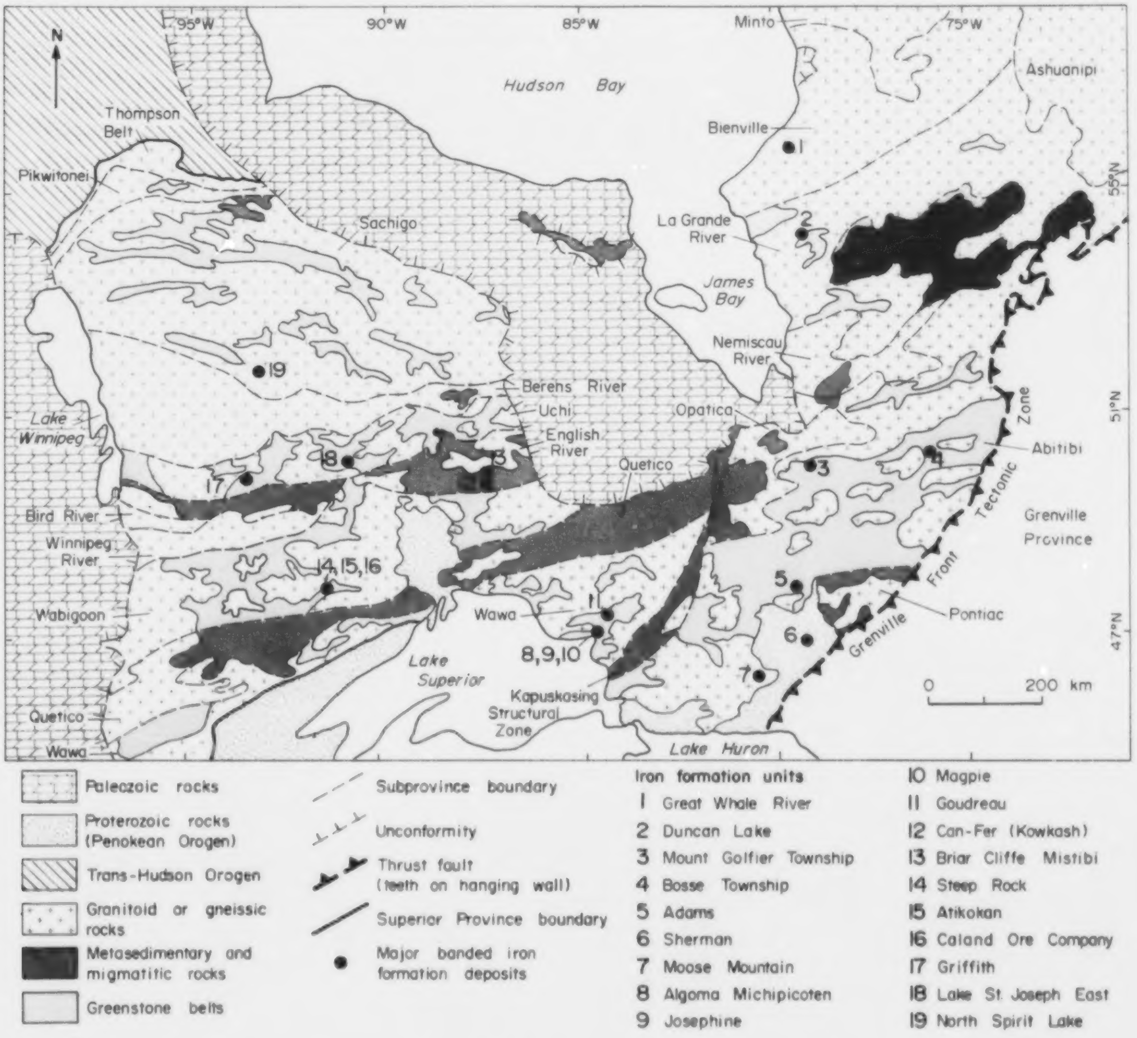

95°W
90°W
85°W
75°W
N
Minto
Hudson Bay
Ashuanipi
Thompson Belt
Bienville
1
55°N
Pikwitonei
La Grande River
2
Sachigo
James Bay
Nemiscau River
19
Berens River
Uchi
English River
Opatica
Zone
51°N
Lake Winnipeg
18
17
3
4
Abitibi
Quetico
Bird River
Tectonic
Grenville Province
Winnipeg River
14,15,16
Front
5
Wabigoon
Wawa
11
Pontiac
47°N
Grenville
6
8,9,10
Lake Superior
7
Kapuskasing Structural Zone
0
200 km
Quetico
Wawa
Lake Huron
Paleozoic rocks
Proterozoic rocks (Penokean Orogen)
Trans-Hudson Orogen
Granitoid or gneissic rocks
Metasedimentary and migmatitic rocks
Greenstone belts
Subprovince boundary
Unconformity
Thrust fault (teeth on hanging wall)
Superior Province boundary
Major banded iron formation deposits
Iron formation units
1 Great Whale River
2 Duncan Lake
3 Mount Golfier Township
4 Bosse Township
5 Adams
6 Sherman
7 Moose Mountain
8 Algoma Michipicoten
9 Josephine
10 Magpie
11 Goudreau
12 Can-Fer (Kowkash)
13 Briar Cliffe Mistibi
14 Steep Rock
15 Atikokan
16 Caland Ore Company
17 Griffith
18 Lake St. Joseph East
19 North Spirit Lake

**Figure 22.10.** Distribution of some major iron formation units in the Superior Province (*after* Franklin and Thorpe 1982).

**Table 22.4.** Depositional settings and ages of major iron formation occurrences in the Superior Province. (Reference given in brackets).

| Setting | Rock Association | Example | Characteristics | Age Estimate (Ga) |
|---|---|---|---|---|
| platform-foredeep (1,2) | quartz-rich clastic sediments, stromatolitic carbonate, iron formation, komatiitic and tholeiitic rocks | a) Disrupted assemblage – North Spirit Lake (3)<br>b) Keeyask assemblage – North Caribou Lake (4)<br>c) Steep Rock Group (5,22) | a) oxide and silicate facies<br>b) oxide and silicate facies<br>c) oxide facies, secondary enrichment | a) ca. 2.9 (6)<br>b) 2.98 (7)<br>c) 2.9 to 3.0 (8) |
| shallow volcanic platform (9); (continental arc?) | tholeiites, rhyolite pyroclastic rocks, carbonate; underlying volcanic rocks altered; oolites preserved in Back River Complex (10) | a) Wawa iron formation in Wawa assemblage (11)<br>b) Pickle Crow assemblage (12) | oolitic texture preserved where undeformed; associated with stromatolites at Wawa (13); laterally equivalent to volcanic-associated, massive sulphide deposits | a) 2.745 (11)<br>b) 2.892 (12) |
| shallow-sea iron formation (9,14) | carbonate, shale, iron formation, tholeiitic sills and flows, felsic volcanic rocks | Archean Hamersley Basin (14). No Superior Province examples known. | oxide and minor silicate facies, shallow sea versus stromatolites, etc. | 2.7 (14) |
| submarine fan (15) | coarse and fine clastic rocks, cross-bedded conglomerates, wackes, iron formation, shale | a) Lake St. Joseph (late unconformable basin) transitional to English River Subprovince (15)<br>b) Namewaminikan Group thrust-based transition from Wabigoon to Quetico subprovinces (16)<br>c) Griffith Mine, English River Subprovince inferred based on tectonic position and rock types to be similar to "b" (17,18) | a) oxide facies, increasingly fine clastic sedimentary component with distance from source<br>b) oxide facies accumulated in inter-channel areas between submarine fans (16)<br>c) oxide-facies iron formation on fold nose | a) 2.713 to 2.71 (12,18)<br>b) 2.72 to 2.69 (11)<br>c) ca. 2.7 (19) |
| mafic plain (2) or starved basin | thick massive tholeiitic-komatiitic flows with thin sulphide-facies iron formation | Jutten Group (20) | sulphide facies with argillite interbeds | >2.74 (19) |
| arc environment (2); deep-water iron formation (9) | a) tholeiites, calc-alkalic andesite-rhyolite<br>b) tholeiitic and komatiitic volcanic rocks, clastic sedimentary rocks;<br>c) tholeiites, calc-alkalic pyroclastic rocks | a) Chambers–Briggs assemblage (21)<br>b) Boston assemblage (21)<br>c) Wabigoon assemblage, Savant Lake area (9,19) | a) oxide and silicate facies<br>b) largely oxide facies<br>c) proximal sulphide facies with graphitic shale, distal oxide facies with turbidites (22) | a) 2.737 (23)<br>b) ca. 2.7<br>c) 2.745 (19) |

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*References: 1– Hoffman 1987; 2– Thurston and Chivers 1990; 3– Wood 1980; 4– Thurston et al., this volume; 5– Wilks and Nisbet 1988; 6– Corfu and Wood 1986; 7– de Kemp 1987; 8– Davis and Jackson 1988; 9– Kimberly 1978; 10– Lambert 1976; 11– Williams et al., this volume; 12– Stott, this volume; 13– Hofmann et al., in press; 14– Trendall 1983; 15– Meyn and Palonen 1980; 16– Barrett and Fralick 1989; 17– Shklanka 1970; 18– Breaks, this volume; 19– Blackburn et al., this volume; 20– Trowell 1986; 21– Jackson and Fyon, this volume; 22– Shegelske 1978; 23– Bowins and Heaman, in press.*

## Metasedimentary Subprovinces

Iron formation in metasedimentary subprovinces accumulated in a submarine fan environment (Meyn and Palonen 1980; Barrett and Fralick 1989). In interfan areas, the iron-rich metasedimentary rocks accumulated as overbank deposits on abandoned lodes (Figures 22.11a and 22.11b; Meyn and Palonen 1980; Barrett and Fralick 1989). The iron formation units tend to occur on the on the oceanic side of the accretionary prism, represented by the sedimentary subprovince.

## Late Unconformable Basins

This environment, also referred to as "Timiskaming-type" basins (Thurston and Chivers 1990), contains relatively minor iron formation units. However, jaspilitic iron formation clasts are common within conglomeratic units in this setting, although similar units are not found in the adjacent "Keewatin-type" greenstone successions (Jensen and Langford 1985), suggesting that iron formation was deposited late in greenstone development and cannibalized during development of late unconformable basins. Alternatively, "Keewatin-type" iron formation was eroded and incorporated into the late unconformable basins.

## Platform-Foredeep

The quartz-rich 2.9 billion-year-old greenstone belts of the North Caribou terrane (Thurston et al., this volume) and the central Wabigoon region (Blackburn et al., this volume) contain oxide- and silicate-facies iron formation (*see* Table 22.4). In the foredeep setting, the sedimentary environment ranges from shallow marine shelves to deep, starved basins (James 1954), with the water deepening with time (Ojakangas 1983). Ample evidence of this is seen in some occurrences in the North Caribou terrane (e.g., Keeyask assemblage, Keewaywin assemblage, Thurston et al., this volume).

# SECULAR VARIATION

Depositional settings of iron formation in the Superior Province (*see* Table 22.4) demonstrate a degree of secular variation, with the platform-foredeep environment concentrated in the 3 to 2.8 Ga interval, whereas the submarine fan occurrences within, and at the margins of, the sedimentary subprovinces tend to be about 2.7 Ga in age.

# ORIGIN

The origin of iron formation is a contentious issue (*see* Holland 1984, p.379 and references therein). There is general consensus that the well-laminated, cherty iron formation formed by chemical precipitation (James and Trendall 1982). The shallow volcanic platform and shallow-sea iron formation types described by Kimberley (1978) may represent deformed and undeformed variants of similar depositional environments. Evidence from undeformed sequences

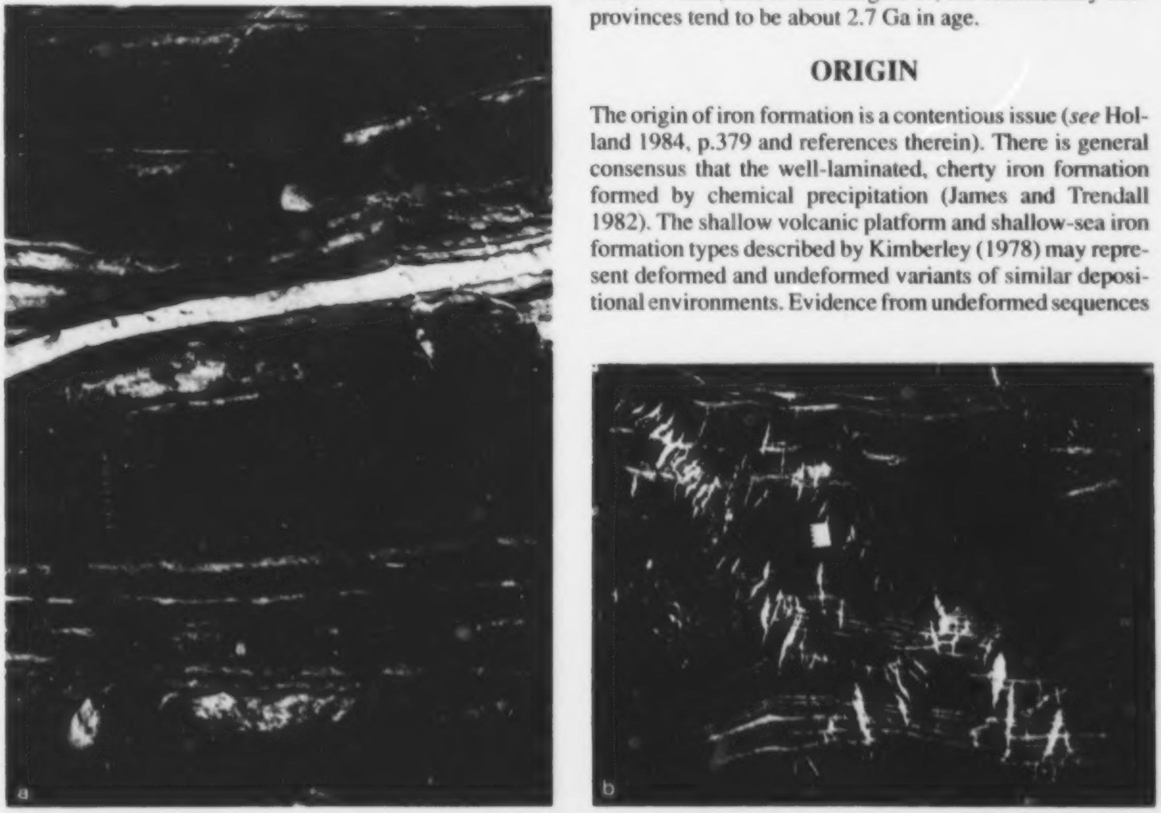

**Figure 22.11.** Oxide-facies iron formation, Sherman Mine, Temagami, that was deposited on a subaqueous fan. **a)** Iron-rich chemical metasedimentary bands that are interlayered with clastic metasedimentary rock, present immediately beneath a thick (greater than 10 m) unit of iron formation (*see* Figure 22.11b). The chert-magnetite mesobands are interlayered with normally graded, feldspathic, volcanic-derived turbidite deposits. The clastic metasedimentary deposits may represent distal turbidite deposits or overbank deposits on a subaqueous fan. **b)** Thick chert, hematite, magnetite iron formation that was a source of iron ore, Sherman Mine, Temagami. This unit occurs stratigraphically above the facies represented in Figure 22.11a. Note the absence of clastic metasedimentary units. The iron formation unit is cut by conjugate shear zones that are defined by arrays of sigmoid, extensional quartz veins.

of Archean age, such as the Hamersley Formation (Trendall 1983), suggest, on the basis of primary structures and biota, deposition in shallow water. The rhythmic iron-rich and iron-poor bands are developed on a millimetre scale, suggesting that this layering was controlled by a diurnal or annual cycle (Holland 1984, p. 402). Therefore, the banding is comparable to varves, in that an annual cycle is inferred. The regional continuity of individual bands in undeformed basins (Trendall 1983), and the grain size of the cherty bands and iron-rich minerals suggests that the layering represents a process of chemical precipitation. Although a direct volcanic source for any of the precipitated components of banded iron formation (BIF) is not demonstrated (James and Trendall 1982), silica and iron may have been indirectly derived by extensive weathering of volcanic materials (Trendall 1983). The iron was transported from source as ferrous iron, which was oxidized by annual algal blooms (Trendall 1983). Origins for iron formation with interbeds of wacke or argillite require cannibalization of unconsolidated iron formation (seen in iron formation clasts in the shallow volcanic type described by Kimberley (1978) at Wawa (Barrett and Fralick 1989)), or deposition in a starved basin (Morganti 1981).

At the peak of Ontario iron production in 1973, 11.3 million tonnes of iron ore were produced annually from Archean iron formations. Many Superior Province iron formations are lean, with grades ranging from 25 to 27% Fe. They are mineable generally only where tectonically thickened, such as the fold noses in the Bruce Lake deposit (Shklanka 1970), or where secondary enrichment has occurred, such as the Steep Rock deposit (Shklanka 1972). Secondary enrichment improved the grade of iron formation from 25 to 35% Fe to in excess of 50% Fe at Steep Rock (Shklanka 1972) and Wawa (Shklanka 1968); the higher grade material is termed "direct-shipping ore". The process of secondary enrichment is generally ascribed to meteoric water (Shklanka 1968; Machado 1987) with the age of kaolinization and secondary enrichment in hematite considered to be Archean (Riley 1970) or Cretaceous (Machado 1987) in age for the Steep Rock deposit.

## INTRUSION-ASSOCIATED METALLIC MINERAL DEPOSITS

Archean plutonism in the Superior Province consisted of the following distinct suites: 1) gneissic tonalite; 2) foliated tonalite; 3) two-mica granite-granodiorite; 4) massive granite-granodiorite; 5) diorite-monzonite-granodiorite; and 6) mafic to ultramafic rock. Early plutonism predated, or was coeval with, volcanism in the volcanic-plutonic subprovinces and included members of the mafic to ultramafic, gneissic tonalite and foliated tonalite suites. These early plutons occur in all subprovinces, and some genetically related metal associations include chromite, nickel-copper-platinum group element, and porphyry copper-molybdenum-(gold).

Late Archean plutonism consisted of two-mica granite-granodiorite, massive granite-granodiorite, foliated tonalite and diorite-monzonite-granodiorite suites. These intrusions occur in volcanic-plutonic, metasedimentary and plutonic subprovinces. Their emplacement postdated most calc-alkalic volcanism in the greenstone belts, but was partly coeval with, and postdated, the alkalic volcanism and associated sedimentation of the "Timiskaming-like" assemblages. Some genetically related metal associations include: 1) various types of pegmatite-associated lithophile elements (rare elements (beryllium, cesium, lithium, niobium, rubidium and tantalum), bismuth, hafnium, molybdenum, tin, thorium, uranium, tungsten and zirconium); and 2) less common mineralization involving certain chalcophile elements, such as platinum group elements, cobalt, copper and nickel.

Pegmatites enriched in certain lithophile elements are dispersed throughout the Superior Province of Ontario (Figure 22.12) and adjacent Manitoba (Cerny and Meintzer 1988). General categories of pegmatite mineralization include (Breaks 1989; Breaks and Janes 1991): 1) rare element pegmatites rich in lithium, rubidium, cesium, beryllium, tantalum, niobium and gallium, which commonly coexist with increased levels of boron, fluorine and phosphorus and typically are depleted in rare earth elements; 2) exomorphic deposits enriched in rare elements, tin tantalum, tungsten and occasional gold as found in albitites, greisens and tourmalinites affiliated with altered rare element pegmatites or their parent plutons; 3) molybdenite-bearing pegmatites accompanied by bismuth, fluorine and possibly trace-level enrichment of some of the rare elements; 4) radio-element-enriched pegmatites which contain uranium and thorium and typically augmented by molybdenum, phosphorus, rare earth elements and zirconium.

The purpose of this section is to provide a brief summary of the plutonic suites, a general review of the related metal associations, and to demonstrate how the plutonism and associated mineralization relates to the tectonic framework of the Superior Province.

### Gneissic and Foliated to Massive Tonalite Suites

The gneissic tonalite suite ranges compositionally from biotite- and hornblende-bearing tonalite to granodiorite (Barrie and Shirey 1991). Older rocks of this suite, dating from 3.2 to 3.0 Ga, occur in the Winnipeg River Subprovince (Corfu 1988; Davis et al. 1988) and parts of the Wabigoon Subprovince (Davis et al. 1988). Younger, *circa* 2.7 billion-year-old, plutonic rocks of the gneissic tonalite suite occur predominantly in volcanic-plutonic subprovinces (Davis et al. 1988).

Plutonic rocks of the foliated tonalite suite constitute a major component of volcanic-plutonic and plutonic subprovinces, and a lesser component of metasedimentary subprovinces (Percival and Williams 1989; Barrie and Shirey 1991). The suite includes foliated and massive tonalite and granodiorite (Barrie and Shirey 1991). Strongly foliated tonalites are common in the plutonic subprovinces and in granitoid terranes external to greenstone belts (Barrie and Shirey 1991). Massive tonalite occurs mainly within volcanic-plutonic subprovinces, in composite gabbro-diorite-tonalite-granodiorite batholiths (Barrie and Shirey 1991). Massive batholiths, such as the Atikwa Batholith in the western Wabigoon Subprovince, may be coeval with calc-alkalic metavolcanic rocks (Davis and Edwards 1986).

## METALLOGENIC ASSOCIATION

No metallic mineral deposits of significance are associated with the gneissic tonalite suite (Sutcliffe and Beakhouse 1990). However, porphyry copper-molybdenum mineralization occurs associated with intrusions of the foliated to massive tonalite-granodiorite suite.

### Porphyry Copper-Molybdenum Mineralization

Few intrusion-associated copper and molybdenum occurrences (*see* Figure 22.12) having characteristics of Phanerozoic porphyry deposits exist in the Superior Province of

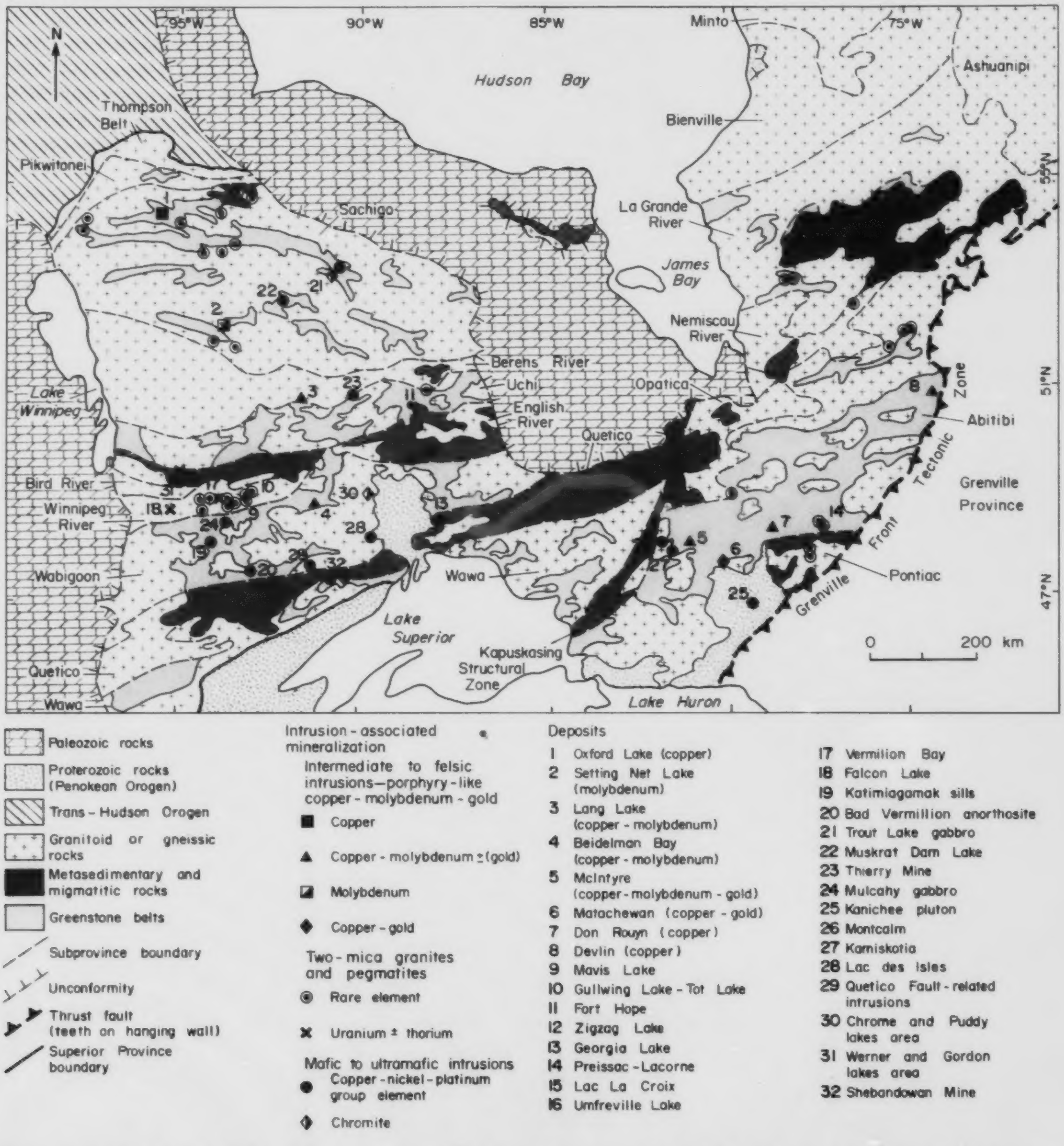
95°W
90°W
85°W
75°W
Minto
Hudson Bay
Ashuanipi
Thompson Belt
Bienville
Pikwitonei
Sachigo
La Grande River
James Bay
Nemiscau River
Berens River
Uchi
English River
Opatica
Quetico
Abitibi
Lake Winnipeg
Bird River
Winnipeg River
Wabigoon
Wawa
Lake Superior
Kapuskasing Structural Zone
Quetico
Wawa
Lake Huron
Grenville Province
Grenville Front Tectonic Zone
Pontiac
55°N
51°N
47°N
0
200 km
N
Paleozoic rocks
Proterozoic rocks (Penokean Orogen)
Trans-Hudson Orogen
Granitoid or gneissic rocks
Metasedimentary and migmatitic rocks
Greenstone belts
Subprovince boundary
Unconformity
Thrust fault (teeth on hanging wall)
Superior Province boundary
Intrusion-associated mineralization
Intermediate to felsic intrusions—porphyry-like copper-molybdenum-gold
■ Copper
▲ Copper-molybdenum ± (gold)
Molybdenum
◆ Copper-gold
Two-mica granites and pegmatites
◉ Rare element
✖ Uranium ± thorium
Mafic to ultramafic intrusions
● Copper-nickel-platinum group element
Chromite
Deposits
1 Oxford Lake (copper)
2 Setting Net Lake (molybdenum)
3 Lang Lake (copper-molybdenum)
4 Beidelman Bay (copper-molybdenum)
5 McIntyre (copper-molybdenum-gold)
6 Matachewan (copper-gold)
7 Don Rouyn (copper)
8 Devlin (copper)
9 Mavis Lake
10 Gullwing Lake-Tot Lake
11 Fort Hope
12 Zigzag Lake
13 Georgia Lake
14 Preissac-Lacorne
15 Lac La Croix
16 Umfreville Lake
17 Vermilion Bay
18 Falcon Lake
19 Katimiagamak sills
20 Bad Vermillion anorthosite
21 Trout Lake gabbro
22 Muskrat Dam Lake
23 Thierry Mine
24 Mulcahy gabbro
25 Kanichee pluton
26 Montcalm
27 Kamiskotia
28 Lac des Isles
29 Quetico Fault-related intrusions
30 Chrome and Puddy lakes area
31 Werner and Gordon lakes area
32 Shebandowan Mine

**Figure 22.12.** Distribution of some intrusion-associated metallic mineral occurrences and deposits in the Superior Province. **a)** Porphyry-like copper-molybdenum-gold occurrences and deposits related to Archean intermediate to felsic intrusions—deposits 1 to 8. **b)** Rare element and other pegmatite-associated fields. Rare element—deposits 9 to 15. Uranium and/or thorium—deposits 16 to 18. **c)** Mafic to ultramafic intrusion-associated nickel-copper-platinum group element and chromite resources. Intrusions comagmatic with associated volcanic rocks—deposits 19 to 27. Late tectonic mafic to ultramafic intrusions—deposits 28 to 32.

**Table 22.5.** Characteristics of some intrusion–associated copper, molybdenum, and gold deposits in the Superior province that display some characteristics of Phanerozoic porphyry deposits (after Ayres and Cerny 1982; Franklin and Thorpe 1982).

| Name (Reference) | Age (Ma) | Pluton Type | Pluton Geometry | Mineralization Metallic Minerals |
|---|---|---|---|---|
| Oxford Lake (2,3,13) | not known | tonalite | | pyrite, chalcopyrite, molybdenite |
| Setting Net Lake (1,11,13,14) | 2712–2710 | granodiorite-granite; potassium feldspar phenocrystic | stock, 5.7 km$^2$ | molybdenite, chalcopyrite, pyrite |
| Lang Lake (4,5,13) | not known | tonalite | sill | chalcopyrite, pyrite, molybdenite, pyrrhotite |
| Beidelman Bay (6,7,13) | 2735 | tonalite with mafic rim | synvolcanic sill | chalcopyrite, pyrrhotite, sphalerite |
| McIntyre (8,9,13) | 2690 | granodiorite | stock | chalcopyrite, pyrite, bornite, molybdenite, tetrahedrite, gold |
| Matachewan (10,13) | ca. 2680 | syenite and quartz syenite | zoned stock | chalcopyrite, pyrite, molybdenite, bornite |
| Don Rouyn (11,13) | ca. 2700 | tonalite, minor granodiorite | synvolcanic stock | inner bornite-chalcopyrite-pyrite zone; middle chalcopyrite-pyrite zone; outer pyrite zone |
| Devlin (12) | ca. 2700 | leucotonalite | Chibougamau Pluton, with magmatic breccia zone | pyrite and chalcopyrite in breccia zone and fractures |

Ontario (Table 22.5; Ayres and Averill 1974; Wolfe 1976; Riley et al. 1971; Kirkham 1972; Colvine and McCarter 1977; Colvine 1978). The McIntyre copper deposit, within the Pearl Lake intrusion, Abitibi greenstone belt (Langford 1941; Pyke and Middleton 1970; Davies and Luhta 1978), is the only deposit that has been mined. Most occurrences are found within, or adjacent to, small subvolcanic to epizonal dikes, stocks and batholiths of tonalite, trondhjemite or granodiorite emplaced into volcanic-plutonic subprovinces (Ayres and Cerny 1982). The paucity of porphyry deposits in the Superior Province (*see* Figure 22.12) is somewhat anomalous, given the presence of tonalite-granodiorite intrusions that were emplaced into cogenetic central volcanic complexes or stratovolcanoes within the greenstone belts.

The copper-molybdenum occurrences generally consist of chalcopyrite and molybdenite in quartz veins. Mineralization consists of stockwork and disseminated chalcopyrite, molybdenite and pyrite occurring in veins and disseminations in fault, fracture, or breccia-controlled zones (Ayres and Cerny 1982). In most of these occurrences, metal zoning consists of a copper-(molybdenum)-rich core and an outer gold-rich zone (Kirkham 1972; Davis and Luhta 1978). Attendant alteration includes albitization, biotitization, sericitization of feldspars and chloritization of ferromagnesian minerals (Franklin and Thorpe 1982). Anhydrite-bearing zones accompany the copper zones at the McIntyre porphyry deposit (Davies and Luhta 1978). The peripheral gold commonly occurs in veins rather than in minor fractures or disseminations (Franklin and Thorpe 1982).

Some of these occurrences, such as the Beidelman Bay occurrence in the Sturgeon Lake area (Poulsen and Franklin 1981; Trowell 1983; Trowell and Johns 1986), are localized in synvolcanic tonalite intrusions that are spatially and temporally associated with VMS mineralization contained within the cogenetic intermediate to felsic metavolcanic successions (Ayres and Cerny 1982). Other porphyry-like copper-molybdenum occurrences, including the McIntyre deposit, are associated with plutons that postdate calc-alkalic volcanism. The McIntyre copper-molybdenum deposit, Abitibi greenstone belt, is associated with the postvolcanic, 2690 million-year-old (Marmont and Corfu 1989), Pearl Lake porphyritic granodiorite (Burrows and Spooner 1986).

## TECTONIC SETTING

Foliated tonalite plutons are associated with gneissic tonalite suite rocks in 3.17 to 2.83 billion-year-old complexes in some volcanic-plutonic and plutonic subprovinces (Barrie and Shirey 1991). Pretectonic, 2.75 to 2.70 billion-year-old

**Table 22.5.** Characteristics of some intrusion–associated copper, molybdenum, and gold deposits in the Superior province that display some characteristics of Phanerozoic porphyry deposits (after Ayres and Cerny 1982; Franklin and Thorpe 1982).

| Mineralization Habit | Deposit Mineralized Area | Deposit Tenor | Alteration |
|---|---|---|---|
| in chlorite-filled fractures | not well characterized | 0.15 to 0.25% Cu, minor molybdenite | epidote, chlorite, pyrite |
| in quartz veins and locally disseminated | 1 $km^2$ | 0.06% $MoS_2$, minor Cu | propylitic (albite, chlorite, sericite, epidote, pyrite, calcite) with incipient phyllic and rare potassic; intensity decreases away from mineralized zone |
| in quartz-calcite-sericite-epidote-chlorite veins and locally disseminated | 850 m x 100 m | 0.4% Cu, minor molybdenite and silver | epidote, sericite, pyrite, calcite, albite, quartz, potassium feldspar, biotite |
| in quartz veins and micro-fractures | 0.3 $km^2$ | 0.06% Cu, 0.03% $MoS_2$ | sericite, chlorite, with biotite, quartz adjacent to chalcopyrite veinlets |
| disseminated and in fractures | ca. 1 $km^2$ | 0.8% Cu; 0.05% Mo; 1 ppm Au (9 x $10^6$ t) | inner albite-quartz-sericite-carbonate; intermediate albite-quartz-anhydrite-hematite; outer quartz-sericite-albite-ankerite |
| disseminated, quartz vein stockwork, fractures | up to 1 $km^2$ | 0.1 to 0.5% Cu, 0.17 to 0.34 ppm Au, minor molybdenite | quartz-chlorite-hematite; potassium feldspar-hematite-biotite-magnetite adjacent to veins |
| disseminated, in fractures, in quartz-ankerite-calcite veins | not well characterized | inner zone >0.25% Cu,<br>middle zone 0.1 to 0.25% Cu,<br>rare molybdenite | hematite in outer zone |
| flat vein, breccia zone, and fractures | 700 m x 1000 m | up to 3% Cu;<br>0.05 to 0.1% Cu | phase I rare quartz-potassium feldspar veins associated with sericite alteration; phase II typical mineralization + quartz-epidote-chlorite veins; phase III hematite veins and replacement |

*References: 1– Ayres and Averill 1974; 2– Elbers 1976; 3– Haskins and Stephenson 1974; 4– Findlay and Ayres 1977; 5– Findlay 1975; 6– Poulsen and Franklin 1981; 7– Davis and Trowell 1982; 8– Burrows and Spooner 1988; 9– Davies and Luhta 1978; 10– Sinclair 1982; 11– Goldie et al. 1979; 12– Guha et al. 1984, 1990; 13– Ayres and Cerny 1982; 14– Corfu and Ayers 1991.*

tonalite-granodiorite intrusions comprise composite batholiths that are spatially and temporally associated with intermediate to felsic metavolcanic successions in volcanic-plutonic subprovinces (Poulsen and Franklin 1981; Trowell and Johns 1986). Foliated tonalite and gneissic tonalite suites are also present as 2.71 to 2.69 billion-year-old complexes that were emplaced into mid-crustal levels (Percival and Krogh 1983) during deformation of volcanic-plutonic subprovinces (Jackson and Sutcliffe 1990). These plutons were derived by partial melting of metasomatized mantle and subducted basalt slab (Capdivila et al. 1982; Martin 1987; Ujike and Goodwin 1987), and derivative melts were emplaced into an extensional arc setting (Barrie and Shirey 1991). Associated porphyry copper-molybdenum mineralization developed in such an extensional setting, within cogenetic central volcanic complexes (Goldie 1976; Gibson and Watkinson 1990).

Postvolcanic plutons associated with porphyry-type mineralization, such as the Pearl Lake granodiorite, were emplaced into plutonic-greenstone subprovinces later during the tectonic history of the Superior Province. These late tonalitic-granodioritic plutons are compositionally similar to those generated, and emplaced, during subduction of oceanic crust and may reflect late, subduction-related magmas (Burrows and Spooner 1989).

## Two-Mica Granite Suite

Muscovite- and/or biotite-bearing, modestly to strongly peraluminous (ASI of Zen (1986) >1.10) plutons of the 2.7 to 2.67 billion-year-old (Percival 1989) two-mica granite-granodiorite suite occur predominantly in the English River and Quetico migmatite metasedimentary subprovinces (Breaks, Cherry et al. 1985; Sawyer 1987; Percival 1989; Williams, this volume; Breaks, this volume) and are considered to be S-type granites. Note that not all intrusions of this suite contain both micas. Many in the English River Subprovince contain biotite only (Breaks, this volume). A few plutons of this suite occur also in the granite-greenstone subprovinces of northwestern Superior Province (Breaks et al. 1986, p.375-378; Breaks and Osmani 1989; Stone 1989, p.27; Stone 1990, p.14), Wabigoon Subprovince (Breaks, Cherry et al. 1985; Sanborn-Barrie

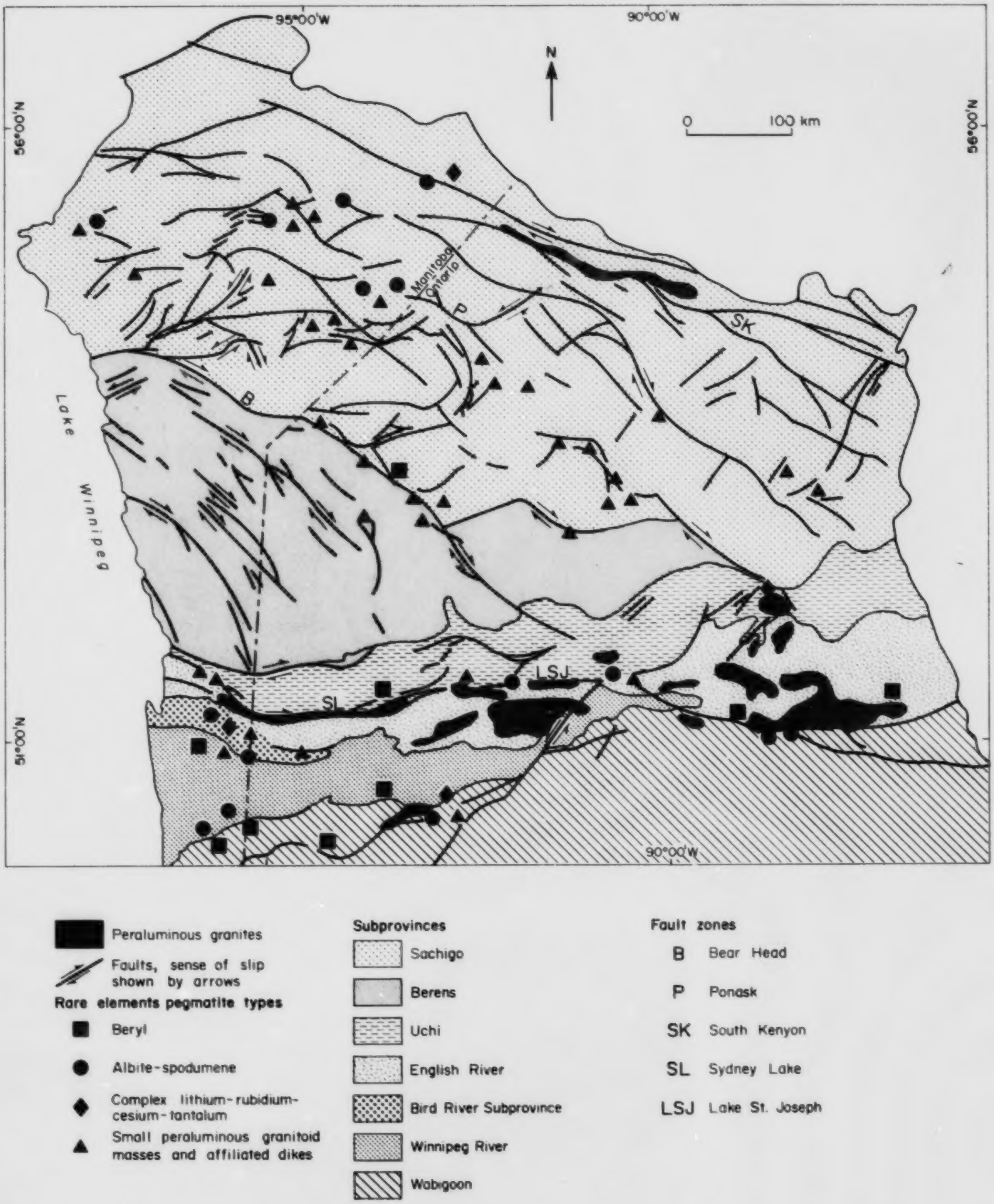
95°00'W
90°00'W
N
0
100 km
56°00'N
51°00'N
Manitoba
Ontario
P
SK
B
Lake Winnipeg
LSJ
SL
90°00'W
Peraluminous granites
Faults, sense of slip shown by arrows
Rare elements pegmatite types
Beryl
Albite-spodumene
Complex lithium-rubidium-cesium-tantalum
Small peraluminous granitoid masses and affiliated dikes
Subprovinces
Sachigo
Berens
Uchi
English River
Bird River Subprovince
Winnipeg River
Wabigoon
Fault zones
B Bear Head
P Ponask
SK South Kenyon
SL Sydney Lake
LSJ Lake St. Joseph

**Figure 22.13.** Distribution of rare metal pegmatites and peraluminous granitoid intrusions in the northwestern part of the Superior Province (Breaks, unpublished data).

1987; Berger 1989), in the North Caribou terrane (Cortis et al. 1988, p.40), Winnipeg River subprovince (Breaks et al. 1978; Breaks, Cherry et al. 1985), and the Abitibi Subprovince (Boily et al. 1990). Very little is known about the distribution of the two-mica granite-granodiorite suite in the high-grade Kapuskasing structural zone or Opatica metasedimentary subprovince.

## METALLOGENIC ASSOCIATION—RARE ELEMENT-BEARING PEGMATITES

Rare element-bearing pegmatites in the Superior Province of Ontario and adjacent Manitoba (*see* Figure 22.12) are spatially associated with peraluminous, plutonic complexes of the two-mica suite (Cerny and Meintzer 1988). This rare element pegmatite mineralization in Ontario occurs (Figure 22.13):

1. within low to medium metamorphic grade zones along terrane boundaries commonly coincident with regional fault systems
   a. between high-grade, metasedimentary and volcanic-plutonic subprovinces (e.g., English River and Uchi subprovinces, English River and Winnipeg River subprovinces, Quetico and Wabigoon subprovinces, and Abitibi and Quetico subprovinces: Cerny and Trueman 1978; Ayres and Cerny 1982; Breaks, Cherry et al. 1985; Breaks 1989)
   b. between high-grade, plutonic and low to medium metamorphic grade, volcanic-plutonic subprovinces (e.g., Winnipeg River and Wabigoon subprovinces: Breaks, Cherry et al. 1985; Stone 1989; Stone 1990)
   c. along late adjustment structures such as the Bear Head fault zone within the North Caribou terrane (Thurston et al., this volume), formerly considered (Card and Ciesielski 1986) as the boundary between the Sachigo and Berens River subprovinces
2. less commonly, within the interiors of Sachigo, English River and Abitibi subprovinces
   a. as deformed bodies within regional fault zones that slice across metavolcanic-metasedimentary belts (e.g., Red Cross Lake belt, Breaks and Osmani 1989)
   b. late tectonic plutons and dikes in supracrustal belts not associated with regional faults (e.g., North Caribou greenstone belt, Breaks et al. 1986 and unpublished data; Preissac–Lacorne Batholith, Boily et al. 1990)
   c. rarely, within medium metamorphic grade internal zones of metasedimentary migmatite subprovinces, such as the English River Subprovince and Quetico Subprovince (Breaks, this volume)
   d. rare pegmatites within the interior of plutonic subprovinces, such as the Winnipeg River Subprovince (Breaks, unpublished data)

The rare element pegmatites constitute one of four major divisions of granitic pegmatites, distinguished on the basis of geological environment and ascribed petrogenesis (Beus 1960; Ginsburg et al. 1979; Cerny 1982, 1989). The rare element class may be divided into several distinct types and subtypes, most of which are found in the Superior Province (Table 22.6; e.g., Fort Hope Field, Hood and Vanstone 1985; Georgia Lake pegmatite field, Zayachivsky 1985, Kissin et al. 1986; Dryden pegmatite field, Breaks 1989). As the Dryden pegmatite field contains a diversity of rare element mineralization and represents one of the most intensely investigated pegmatite fields in Ontario in terms of its geological, petrographic, and chemical features, it is described in more detail.

### Dryden Pegmatite Field

The 10 by 30 km Dryden pegmatite field (Mulligan 1965) is situated within the boundary zone, termed the Sioux Lookout domain (Beakhouse 1988 and 1989), between the Winnipeg River Subprovince and the Wabigoon Subprovince. The pegmatite field is divided into 2 areally distinct pegmatite populations (Breaks, Cherry et al. 1985): 1) Mavis Lake pegmatite group; and 2) Gullwing Lake–Tot Lake pegmatite group. Rare element pegmatites of the Mavis Lake pegmatite group are spatially and genetically linked with the adjacent, peraluminous S-type Ghost Lake Pluton (Breaks 1989, 1990). This late tectonic, multistage, comagmatic, subsolvus, 280 km$^2$ complex was emplaced principally into medium and high metamorphic grade clastic metasedimentary rocks. However, the intrusion cuts across, and hence postdates, the regional sillimanite-muscovite isograd at the boundary between the Sioux Lookout domain (Beakhouse 1988) and the Wabigoon Subprovince (Breaks 1989). Rock types in the Ghost Lake Batholith include: 1) biotite and cordierite-biotite granite; 2) muscovite-biotite granite; 3) granodiorite; and 4) a small zone (10 km$^2$) of pegmatitic granite interpreted as a cupola zone (Cerny and Meintzer 1988). The pegmatitic granite phase occupies the eastern lobe of the intrusive complex. Notable accessory and varietal minerals in the Ghost Lake Pluton include tourmaline, garnet, sillimanite, and scarce apatite, monazite, zircon, dumortierite, beryl and tapiolite (Breaks 1989; Breaks and Janes 1991). The inferred cupola zone has the bulk chemical characteristics and chemical indices comparable to fertile pegmatitic granite masses in other fields (Cerny and Meintzer 1988).

### Mavis Lake Pegmatite Group

The Mavis Lake pegmatite group occurs as a 6 km long endo- and exocontact aureole associated with the Ghost Lake Batholith. With increasing distance east from the batholith, the field exhibits the following regional zonation of pegmatite types and subtypes (Breaks 1989; Breaks and Janes 1991):

1. endogenous beryl [Be-B-(Cs) and Rb-Be-F-Sn-(Cs-Ga-Ta>Nb)]
2. beryl-columbite [Be-B-Nb>Ta-P-(Cs)]
3. albite-spodumene [Li-Rb-Be-Ta>Nb-B]
4. albite-type [Li>Rb-Be-Ta>Nb and Rb>Li-Be-Ta>Nb].

Table 22.6. Classification of rare–element pegmatites of the Superior province (after Cerny 1989).

| Pegmatite Type | Subtype | Mineralogy | Metal Association | Association Examples | Reference[1] |
|---|---|---|---|---|---|
| beryl | a) beryl-columbite<br>b) beryl-columbite-phosphate | a) beryl, tourmaline, columbite-tantalite, (apatite)[2], (garnet)<br>b) beryl, columbite-tantalite, purpurite, apatite, tourmaline | a) Be-B-(Nb>Ta-P)[2]<br>b) Be-P-B-(Nb>Ta) | a) Taylor pegmatites Nos. 1 and 2<br>b) Sandy Creek pegmatites | a) (1)<br>b) (2) |
| albite-spodumene | | green spodumene, beryl, tantalite, columbite, (elbaite), (spessartine) | Li-Rb-(Be-Ta>Nb-P) | a) Fairservice pegmatites (north, northeast and south zone)<br>b) Root Lake Field | a) (1,3)<br>b) (3,4) |
| complex type | a) spodumene<br>b) petalite<br>c) amblygonite<br>d) lepidolite | a) pink spodumene, pollucite, beryl, tantalite, columbite, (elbaite), (spessartine)<br>b) spodumene-quartz intergrowth, purpurite, columbite, amblygonite, beryl, (cassiterite), (hühnerkobelite)<br>c) amblygonite, beryl, tantalite, (lepidolite), (pollucite)<br>d) lepidolite, microlite | a) Li-Rb-Cs-(Be-Ta>Nb)<br>b) Li-P-F-(B-Be-Ta=Nb)<br>c) F-Li-Rb-Cs-Be-Ta>Nb-(Sn-B-Ga)<br>d) Li-Ta>Nb-F | a) Tot Lake pegmatite, Lily Pad Lake pegmatite<br>b) MNW pegmatite<br>c) unknown in Ontario<br>d) rare in Ontario and only known as small secondary pods in Gullwing Lake pegmatite | a) (1,3,5,6,7)<br>b) (8,9,10,11)<br>c) (12)<br>d) (1) |
| albite | | beryl, tantalite | Be-Ta>Nb | Fairservice pegmatites (East zone) | (1) |

[1] *References: 1– Breaks 1989; 2– Breaks, unpublished data; 3– Mulligan 1965; 4– Breaks et al. 1978; 5– Prest 1942; 6– Wallace 1978; 7– Hood and Vanstone 1985; 8– Milne 1964; 9– Pye 1965; 10– Breaks 1980; 11– Zayachivsky 1985; 12– Cerny 1991.*

[2] *Brackets indicate trace abundances of elements or mineral phases.*

This zonal hierarchy, similar to that found in the Beaulieu District in the Slave Province (Hutchinson 1955), is also sympathetic with increasing extent of albitization and greisenization (Breaks 1989).

## Gullwing Lake–Tot Lake Pegmatite Group

The 1 to 2 by 15 km Gullwing Lake–Tot Lake pegmatite cluster partially envelopes the syntectonic Lateral Lake stock of the foliated tonalite suite. The cluster occurs mainly within highly deformed amphibolitic mafic metavolcanic rocks, minor clastic metasedimentary rocks (<2698±4 Ma, Davis et al.1988) and tonalites of the Lateral Lake stock. The Gullwing Lake–Tot Lake pegmatite group differs from the Mavis Lake pegmatite group in the following respects: 1) lack of a well-defined regional zonation of pegmatite types; 2) diverse metal association, varying from [Mo-Bi-Cu] to [Li-Cs-Rb-Be-F-Ta>Nb]; 3) presence of a complex rare element pegmatite of the spodumene subtype which exhibits the uncommon coexistence with pollucite; 4) absence of extensive boron-rich exomorphism and associated scheelite; and 5) absence of a clearly defined parent granite pluton. The Gullwing Lake spodumene-columbite-beryl-molybdenite pegmatite is notable in that it demonstrates, within a single intrusive dike, a genetic linkage between a primary lithium aluminosilicate-bearing core zone and a pegmatitic granite unit. The latter is petrographically similar to potassic pegmatite units in chemically evolved parts of the Ghost Lake Pluton. Physical linkage with the Ghost Lake Pluton, situated 10 km to the southwest, is uncertain. This affiliation has only rarely been noted elsewhere in Archean sequences, including the petalite-bearing pegmatitic leucogranite complex at Red Sucker Lake, Manitoba (Chackowsky and Cerny 1986).

The Tot Lake pegmatite (Figure 22.14), belonging to the complex spodumene subtype, is characterized by chemical fractionation indices that closely approach the Tanco Pegmatite (Cerny et al. 1981; Breaks 1989). Both of these pegmatites contain pollucite, a mineral whose presence is indicative of extreme fractionation (Cerny and Meintzer 1988).

## Preissac–Lacorne Batholith

Most of the lithium-, beryllium-, tantalum- and molybdenum-bearing pegmatites in the Abitibi Subprovince occur in the Preissac–Lacorne Batholith in Quebec. The geology of the batholith is described here to offer some guidance in the search for similar deposits in the western part of the Abitibi Subprovince. The batholith consists of (Dawson 1966; Boily et al. 1990): 1) early metaluminous, foliated cumulate gabbro, diorite, monzodiorite, monzonite and granodiorite; 2) late foliated to massive, peraluminous monzogranitic plutons (Lacorne, La Motte and Preissac) surrounded by an aureole of rare element- and molybdenite-bearing pegmatites. The Preissac monzogranite has a Pb-Pb age of 2690±40 Ma (Gariépy and Allègre 1985) and a U-Pb zircon age of 2655 Ma (Steiger and Wasserburg 1969). Molybdenum mineralization occurs mainly in quartz-muscovite-potassium feldspar veins spatially and temporally related to the muscovite-bearing monzogranites (Boily et al. 1990).

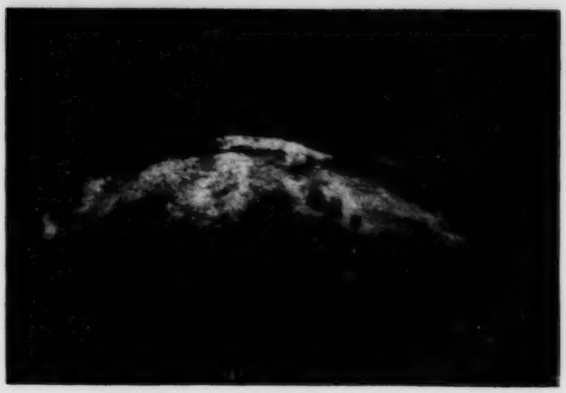

**Figure 22.14.** Longitudinal layering developed in the Tot Lake rare element pegmatite. A late, central quartz vein (Q) cuts a spodumene-rich zone (S) and finer-grained albite unit (A). P. Cerny stands on a pollucite zone. (Photo *courtesy of* F.W. Breaks, Ontario Geological Survey).

Deposits of lithium, beryllium and tantalum are restricted to pegmatitic dikes and pods within monzogranite or an exomorphic aureole (Boily et al. 1990).

The Lacorne, La Motte and Preissac plutons consist of biotite and biotite-muscovite monzogranite with rare garnetiferous variants. The plutons and the surrounding units are cut by pegmatite dikes with the following element associations: 1) beryllium and tantalum with a fringe of lithium mineralization; 2) lithium, beryllium and tantalum associated with molybdenum-bearing veins; and 3) molybdenum-bearing veins (Boily et al. 1990).

The lithium-, beryllium- and tantalum-bearing pegmatite dikes are spatially zoned about the parent monzogranites. Spodumene-rich pegmatites are generally unzoned or poorly zoned. They occur within the surrounding supracrustal and intrusive rocks (Boily et al. 1990). The tantalum- and beryllium-bearing pegmatite dikes occur towards the margins of the parent monzogranite plutons (Boily et al. 1990).

The Case batholith is another two-mica granitic intrusion in the western part of the Abitibi greenstone belt. This intrusion is located in Ontario along the contact between the Abitibi greenstone belt and the Opatica metasedimentary subprovince, northwest of the Abitibi batholith (e.g., Jackson and Fyon, this volume). Spatially associated with the intrusion are spodumene, columbite-tantalite and molybdenite-bearing pegmatite dikes (Lumbers 1962).

## MOLYBDENITE-RICH PEGMATITES

The eastern extremity of the Gullwing Lake–Tot Lake pegmatite swarm is a cluster of molybdenite-rich, potassic pegmatites which includes the Pidgeon molybdenite deposit (*see* Figure 22.12; Colvine and McCarter 1977; McCarter 1980; Page 1984). The Pidgeon deposit is one of the largest molybdenum-bearing pegmatites in northwestern Ontario, containing a resource of 15.8 million tons at 0.08 % $MoS_2$ (Janes and Huggins 1982). $MoS_2$ mineralization has been attributed to an hydrothermal system developed during the late evolution of the Lateral Lake stock (Colvine and McCarter 1977; McCarter 1980). Alternatively, Page (1984) concluded that the $MoS_2$ pegmatites are posttectonic with respect to deformation in the stock and suggests that pegmatite emplacement was governed by tension fractures developed by displacement along the nearby Kathlyn Lake Fault.

The timing of these more primitive pegmatites relative to rare element pegmatites of the Gullwing Lake–Tot Lake pegmatite group is uncertain. However, a genetic association is inferred, based on the presence of accessory molybdenite in most of the rare element pegmatites, the sporadic appearance of rare element minerals, beryl and samarskite, in some of the molybdenite pegmatites, and elevated abundances of several rare elements (Li, Cs, Rb and Ga), Sn and W in microcline and muscovite in the potassic pegmatites at the Pidgeon molybdenite deposit (Breaks and Janes 1991).

Molybdenite-bearing pegmatites are also spatially and temporally associated with the Preissac pluton in the Preissac-Lacorne Batholith (*see* Preissac-Lacorne Batholith, above).

## RARE ELEMENT-, TIN- AND TUNGSTEN-BEARING DEPOSITS

Elevated concentrations of rare elements, tin and tungsten occur in some rocks of the Superior Province that have undergone metasomatic transformation due to interaction with pegmatite- and/or pegmatitic granite-derived fluids (Breaks 1988, 1989). The altered rocks form zones of albitization or greisenization (Breaks 1988, 1989). These alteration zones are either controlled by the contact of fertile pegmatitic granites with their supracrustal host or by shear zones that served as fluid conduits (Breaks and Janes 1991). These deposits have not been widely recognized in the Superior Province and the best example is the endogenous beryl zone of the Mavis Lake pegmatite group (Breaks 1989; Breaks and Janes 1991). This type of exomorphic mineralization may also contain cesium (Ovchinnikov et al. 1972).

Several types of element enrichment have been identified in exomorphic zones (Breaks 1989): 1) synmagmatic holmquistite-actinolite-phlogopite-dravite-bearing shear zones, enriched in boron, lithium, fluorine, rubidium, cesium and tin that may be unique in the Superior Province and similar to lithium-rubidium-cesium metasomatic rocks in the USSR (Gordiyenko et al. 1975); 2) postmagmatic, zones, enriched in potassium, rubidium, lithium, cesium and fluorine, of marginal albitization in pegmatitic granites and complementary metasomatic selvages in adjacent amphibolitic host rocks; 3) exocontact beryl-cassiterite, mica-rich greisen (Kuhne et al. 1972) associated with cassiterite-beryl-tourmaline albitites; and 4) scheelite-bearing calc-silicate pods and tourmalinite sheets.

Two habits of tungsten mineralization occur within the Mavis Lake pegmatite group (*see* Figure 22.12). Minor amounts of scheelite occur in quartz-calcite-diopside-epidote-garnet-vesuvianite-axinite calc-silicate pods and layers within deformed, amphibolitized, mafic metavolcanic flows (Breaks, Cherry et al. 1985). Higher abundances of scheelite occur in biotite-, tourmaline- and sulphide-rich zones up to 0.3 m thick, accompanied by quartz, plagioclase, pyrite and rare holmquistite (Breaks, Cherry et al. 1985).

Cassiterite-, muscovite-, garnet-, magnetite-, and topaz-bearing pegmatite dikes cut a biotite- and muscovite- bearing granite in the Linklater Lake area, northwest of Lake Nipigon (Chisholm 1948). This occurrence was the first discovery of tin in Ontario. Beryl-, tourmaline- and garnet-bearing pegmatitic dikes also occur in the area, but contain no cassiterite (Chisholm 1948). The pegmatites are deformed by shearing related to the Pashkokogan Fault Zone. The segment along which the pegmatites and spatially associated two-mica granite occur is a boundary fault that separates the Wabigoon from the English River subprovinces.

## URANIUM- AND/OR THORIUM-RICH PEGMATITES

This class of granitic pegmatite, abundant in the Vermilion Bay and Umfreville Lake pegmatite fields (*see* Figure 22.12), most closely compares with the Rare Earth Type described by Cerny (1989), although chemical data on the Ontario representatives are sparse (Breaks 1982). This pegmatite class is associated with intrusions of the peraluminous (one- and two-mica) granite-granodiorite suite and (Type 1) and massive granite-granodiorite suite (Type 2) (Breaks 1982).

Type 1 uranium-thorium-bearing pegmatites are associated with coarse-grained, S-type, pegmatitic biotite granite and granodiorite (Breaks 1982). These pegmatites, found mainly in the English River Subprovince, constitute most of the Umfreville Lake pegmatite field (*see* Figure 22.12). Pegmatites are located about the periphery of the Umfreville–Conifer lakes granulite centre (Breaks 1982). Secondary muscovite, after biotite, may be present in uranium-thorium pegmatites confined to ductile deformation zones, such as the Sydney Lake occurrence (Breaks et al. 1978; Stone 1981; Breaks 1982). This type of pegmatite may be derived by anatexis of an interbedded pelite and metawacke precursor (Breaks 1982).

Type 2 uranium-thorium pegmatites are affiliated with metaluminous to weakly peraluminous (ASI = 0.98 to 1.10) biotite granite, granodiorite and rare syenite (Breaks 1982) of probable I-type genesis (Beakhouse 1983; Beakhouse and McNutt, 1991). These pegmatites are mainly concentrated in the Vermilion Bay pegmatite field, along the Winnipeg River Subprovince–Wabigoon Subprovince boundary zone. The potassic pegmatites occur as internal segregations in the outer confines of, and were largely derived from, the late tectonic to posttectonic Feist Lake stock (Pryslak 1976; Breaks 1982) which forms part of the Dryberry Batholith (Sanborn-Barrie 1987). Pegmatites, such as the Richard Lake deposit and the Hawk Lake occurrence, also occur as exocontact masses within amphibolitic mafic metavolcanic rocks of the Tustin–Bridges greenstone belt (Pryslak 1976; Breaks and Janes 1991).

## DEPOSITIONAL MODEL

Several factors play an important role in the generation and localization of rare element and uranium-thorium pegmatite mineralization. The rare element pegmatites are genetically linked with chemically evolved two-mica granitoids in parts of the Superior Province (e.g., Dryden pegmatite field, Breaks 1989; Osis Lake area, Birse Lake pegmatite swarm, Cerny et al. 1981). These granites were emplaced into medium-grade metamorphic domains near, and along, the contact with adjacent granite-greenstone terranes. In these areas, a metamorphic gradient from high to medium grade is best developed, and migmatitic metasedimentary rocks are invariably present (Breaks, Cherry et al. 1985). An S-type derivation is inferred for two-mica granitoids that are distributed within the low- and medium-grade transition zones between migmatitic metasedimentary and volcanic-plutonic subprovinces. The pegmatitic granites represent the products of extensive chemical fractionation (e.g., Greer Lake stock, Cerny et al. 1981). Where the parental granite is not exposed, such as at Bernic Lake (Cerny et al. 1981), a linkage to two-mica granitoids is less certain, and other petrogenetic mechanisms may be considered in the pegmatite generation (Zen 1986, 1988).

The confinement of rare element pegmatites to low- and medium-grade domains of the high temperature-low pressure Abukuma-type of regional metamorphism may be attributed to the following factors: 1) occurrence of "pre-intrusion strength anisotropies" (Brisbin 1986, p. 649), such as joints and cleavages, especially in mafic metavolcanic rocks which host the majority of Superior Province rare element pegmatites; 2) presence of major shear zones, guided by subprovince boundaries and granite-greenstone contacts within certain subprovinces (Sachigo; Breaks and Osmani 1989) and their role in the dispersion of pegmatite melt-fluid systems (Beus 1962, 1968; Cerny 1989); 3) influence of volatile components (boron, fluorine, $H_2O$) and lithium and $P_2O_5$ on liquidus and solidus temperatures, and on major, minor and trace element partition coefficients (Wyllie and Tuttle 1961 1964; Koster van Groos and Wyllie 1968; Bankwitz 1974; Chorlton and Martin 1978; Manning 1981; Pichavant 1981, 1984; Benard et al. 1985; London 1986). The presence of hyperfusible components (boron, fluorine, $H_2O$, lithium and $P_2O_5$) are responsible for the extended fractionation history and depolymerization (Burnham 1979) of magmas. The resultant decrease in magma solidus and liquidus temperatures and viscosity increased the mobility of fertile granite and rare element pegmatite melt-fluid systems (Bankwitz 1974; London 1986). These fertile magma systems attained solidus temperatures as low as 420° to 470°C for the Tanco Pegmatite (London 1986) and 470°C for pegmatites of the albite-spodumene type in the Dryden pegmatite field (Taylor et al. 1991). This is approximately 200°C lower than that of peraluminous granite melts which are essentially barren of hyperfusible components (Huang and Wyllie 1973). Consequently, the fertile granite-rare-element pegmatite residual melt-fluid systems remained mobile long after the consolidation of their chemically primitive predecessors and were able to access the lower-temperature, brittle-fracture domains along the low to medium metamorphic grade subprovince boundary zones. Conversely, the less fractionated parental biotite and cordierite-biotite peraluminous granite plutons are barren of rare element mineralization and are commonly associated with granulite grade metasedimentary migmatites in the interior of the English River Subprovince.

Uranium-thorium-bearing pegmatites of the S-type, anatectic, peraluminous granite affinity do not contain significant hyperfusible components and are chemically more primitive than those associated with the rare element pegmatites (Breaks, Cherry et al. 1985). These pegmatites are largely confined to high-grade metamorphic parts of the English River Subprovince. Pegmatite mineralization and emplacement was more dependent upon uranium-thorium-enriched metasedimentary protoliths, mobilization of uranium and thorium during regional metamorphism to the periphery of low-pressure granulite centres, and possibly to local structural settings. Local structural settings controlled the emplacement of uranium-thorium pegmatites in the English River Subprovince in highly deformed, peraluminous granite along the Sydney Lake–Lake St. Joseph Fault (Breaks 1989). The shear heating and hydration model (Strong and Hanmer 1981) may have applicability to the mobilization of radio elements within this fault system.

## TEMPORAL CONSTRAINTS

Medium- and high-grade metamorphism in the English River Subprovince and Winnipeg River Subprovince was integral to the formation and localization of metalliferous granitic pegmatites. High-grade metamorphism in the Winnipeg River Subprovince, locally elevated to low-pressure granulite conditions, took place in the Kenora area around 2709 Ma, and at 2680 Ma in the Daniels–Cedar lakes area to the east (Corfu 1988). Emplacement of most late to posttectonic plutons occurred between 2705 and 2660 Ma (Beakhouse, this volume; Corfu 1988). In the English River Subprovince, peraluminous one- and two- mica leucosomes in metasedimentary migmatites were produced by high-grade to granulite-facies regional metamorphism between 2692 and 2668 Ma (Krogh et al. 1976; Corfu, unpublished data). Rare element- and uranium-thorium-bearing granitic melts are inferred to have been initiated during this interval. However, the emplacement of highly fractionated, mobile, fertile pegmatitic granite and related rare element pegmatites continued after most tectonism.

## TECTONIC SETTING

The tectonic setting of rare element, molybdenum- and uranium-thorium-bearing pegmatites is defined by the setting in which the various genetically related granitic suites were generated and emplaced. Many of these pegmatite types are directly affiliated with the peraluminous two-mica granite-granodiorite suite (Breaks, Cherry et al. 1985). These rocks can be produced in a variety of tectonic regimes under different partial melting conditions affecting a variety of sources (Pitcher 1982). Copious volumes of peraluminous, two-mica granitic melt can result from fluid-absent anatexis of compositionally favourable metawacke and metamudstone induced by dehydration of biotite under granulite conditions (Vielzeuf and Holloway 1988). Evidence supporting this mechanism is found in the English River Subprovince, where a spatial relationship exists between granulite centres and the largest masses of the peraluminous two-mica granite-granodiorite (Breaks, this volume).

The petrogenesis of peraluminous granitoids, principally of Phanerozoic age, has been ascribed to a variety of tectonic settings:

1. collisional environments
    a. continent-continent (Le Fort 1981)
    b. arc-continent (Beakhouse et al. 1989)
2. transcurrent fault systems of regional extent (Strong and Hanmer 1981)
3. accretionary prisms (Hill et al. 1981)
4. closing, sediment-dominated, back-arc basin (Lalonde 1989)
5. intracontinental rift systems (Sandiford and Powell 1986; Clemens and Vielzeuf 1987)

Peraluminous granites are widespread in collisional environments and may have been generated by fluid-present anatexis of quartzofeldspathic rocks (e.g., clastic metasedimentary rocks, siliceous tonalites) induced by hydration at the base of thrust sheets (Le Fort 1981; Harris et al. 1986). In Phanerozoic terranes, these syntectonic to late tectonic Hercino-type (Pitcher 1982) granitoids are commonly associated with tin, tungsten and rare element mineralization (Sawkins 1990).

Although the petrogenesis of two-mica granitoids is a characteristic of continent-continent collisional orogenic belts (Le Fort 1981), such intrusions can also develop within an intracontinental rift setting (Clemens and Vielzeuf 1987) or during metamorphism in continental arc-type environments (Bohlen 1987). Within an intracontinental rift setting, anatexis necessary to generate peraluminous magmatism can occur at relatively low pressures (0.3 to 0.7 Gpa) in response to crustal thinning and elevated heat flow (Sandiford and Powell 1986).

Of these tectonic settings, the geometrical relationship between the Wawa, Wabigoon, Winnipeg River, English River and Uchi subprovinces in the western Superior Province is attributed to a collisional event (Williams, Stott, Thurston et al., this volume). Thus, we attribute the large proportion of peraluminous magmatism to collision with, and possible augmentation from, accretionary prisms in metasediment-dominated subprovinces (Percival and Williams 1989; Breaks, this volume; Thurston et al., this volume; Williams, Stott, Thurston et al., this volume).

Along the boundaries between volcanic-plutonic and plutonic subprovinces, such as between the Sachigo and Berens River subprovinces, peraluminous granites occur in regional shear zones, such as the Bear Head fault zone (Stone 1989, Stone 1990). These plutons occur as elongate bodies mostly within the fault zone. They were possibly generated by anatexis facilitated by ingress of volatiles and frictional heating (Breaks, this volume) during a collisional event.

In volcanic-plutonic or plutonic subprovinces, fractional crystallization of I-type tonalitic suites may have resulted in elimination of metaluminous phases such as hornblende and pyroxene, and can shift residual magmas towards peraluminous compositions (Cawthorn and Brown 1976; Cawthorn et al. 1976). However, the contribution of such peraluminous two-mica granites of I-type affinity has yet to be recognized in the Superior Province of Ontario.

## Granite-Granodiorite Suite

Massive, commonly microcline-megacrystic, biotite granite and granodiorite plutons, such as the White Otter Lake batholith (Schwerdtner et al. 1979) in the Wabigoon Subprovince and the Lount Lake batholith in the Winnipeg River Subprovince, are an important, relatively late component of the volcanic-plutonic and plutonic subprovinces (Barrie and Shirey 1991). The Abitibi Batholith in the Abitibi Subprovince (Mortensen 1987) and the Lount Lake batholith (Beakhouse and McNutt, 1991) postdate 2.7 Ga.

### METALLOGENIC ASSOCIATION

Whereas most rare element mineralization in pegmatites is associated with granodiorite and granite of the two-mica suite, Breaks (1989 and unpublished data) has documented beryl and samarskite in molybdenite-bearing potassic pegmatite and yttrotantalite-bearing pegmatite dykes, both of which have an affinity to the massive granite-granodiorite suite. Cerny et al. (1981) have described rare element pegmatites of the Rare Earth Type which are genetically related to the Lac du Bonnet Pluton in southeastern Manitoba, a member of the massive granite-granodiorite suite. This mineralization, which has some A-Type magmatic characteristics (Cerny et al. 1987), may potentially be present in the Winnipeg River Subprovince of Ontario, inferred from the widespread distribution of chemically identical granitoid intrusions (Breaks et al. 1978; Beakhouse, this volume) and by the sporadic appearance of yttrotantalite north of Kenora (Breaks, unpublished data).

Uranium-thorium-bearing pegmatites associated with the massive granite-granodiorite suite are late to post-tectonic. In the English River Subprovince and in the Dryden area, emplacement of this suite overlaps that of the two-mica peraluminous granite-granodiorite suite (Breaks 1989, this volume). However, specific temporal relations between the uranium-thorium-bearing, coarse-grained to pegmatitic units associated with both suites remains uncertain. The abundance of uranium-thorium-bearing pegmatites in areas devoid of metasedimentary migmatites implies that the direct genetic affiliation between Type 2 uranium-thorium mineralization and plutons of the I-Type massive granite-granodiorite suite bears no relationship with the metamorphic history of the English River Subprovince (Breaks 1982). In those areas where Type 2 uranium-thorium pegmatites occur in proximity to the English River Subprovince (e.g., Umfreville Lake pegmatite field), pegmatite formation may have been enhanced by contamination of I-Type magmas with uranium-thorium-enriched metasedimentary rocks.

A direct genetic relationship between these uraniferous pegmatites and plutons of this suite may not be related to the metamorphic history of metasedimentary rocks. However, plutons of the granite-granodiorite suite are associated with uraniferous pegmatites only where the plutons occur in, or proximal to, metasedimentary domains. Hence, the uraniferous pegmatite-granite suite association may implicate: 1) interaction between the metasedimentary rocks and an evolving potassic magma during the magma emplacement in and passage through the metasedimentary package; and 2) subsequent fractionation history of the magma.

### TECTONIC SETTING

Plutons of the granite-granodiorite suite in the Winnipeg River Subprovince, such as the Lount Lake batholith, formed by partial melting of a tonalite protolith induced by high-grade metamorphism in a tectonically thickened crust (Beakhouse and McNutt 1991). The suite is also present in the granite-greenstone subprovinces, where they are chemically akin to the foliated tonalite suite (Barrie and Shirey 1991). A continental arc may be an analogous tectonic setting (Barrie and Shirey 1991).

## Diorite-Monzonite and Diorite-Syenite Suites

Zoned diorite-monzodiorite-monzonite-granodiorite plutons (Stern et al. 1989) are common in volcanic-plutonic subprovinces, but occur in all subprovince types (Barrie and Shirey 1991). The plutons were intruded (Barrie and Shirey 1991): 1) into the metavolcanic successions; 2) into tonalite suites; and 3) along the contact between earlier tonalite and metavolcanic successions in composite granitoid complexes. These plutons are generally late tectonic and less than 2.71 billion years old in the Wabigoon Subprovince (Davis et al. 1988) and less than 2.70 billion years old in the Abitibi Subprovince (Corfu et al. 1989). The suite is associated with amphibole- or biotite-rich, mafic lamprophyre dikes and amphibole- or pyroxene-rich intrusions (Sutcliffe et al. 1990).

The late tectonic, diorite-syenite suite consists of silica over- to under-saturated plutons (Barrie and Shirey 1991). In the Abitibi Subprovince, the 2.68 to 2.66 billion-year-old (Corfu et al. 1989) diorite-syenite intrusions are spatially and temporally associated with Timiskaming-type volcanic and sedimentary successions (Cooke and Moorhouse 1969).

### METALLOGENIC ASSOCIATIONS

#### Lode Gold

Plutons of the diorite-monzonite and diorite-syenite suites are spatially associated with lode gold mineralization in the Kirkland Lake area of the Abitibi greenstone belt (Cooke and Moorhouse 1969). Although a genetic relationship is postulated (Cameron and Carrigan 1987; Cameron and Hattori 1987), mineralization appears to be at least

30 million years younger (Masliwec et al. 1986; Hanes et al. 1987; Wong et al. 1989; Jemielita et al. 1990), precluding a direct genetic relationship with the syenites.

### Alkalic Intrusion-Associated Copper-Gold-Molybdenum Mineralization

Porphyry-like copper-gold-molybdenum mineralization, such as the Ryan Lake and Stancop deposits in the Abitibi greenstone belt, are associated with *circa* 2.68 billion-year-old syenites, principally in the Matachewan area (Sinclair 1979, 1982). Mineralization consists of chalcopyrite, molybdenite and gold in quartz vein stockworks, fractures and disseminations (Sinclair 1979). Associated alteration includes hematitization, potassium metasomatism, silicification and chloritization (Sinclair 1979). The proximity of these deposits to the western part of the Larder-Cadillac fault zone and attendant lode gold mineralization complicates the classification of this copper-gold-molybdenum mineralization in the Matachewan area.

Syenite intrusion-associated gold mineralization also occurs in the Shebandowan greenstone belt, within, and adjacent to, Timiskaming-like successions. At the Stewart occurrence in Conmee Township (Carter 1990), gold, traces of chalcopyrite and up to 20% pyrite occur within an arcuate zone that consists of silicified and sheared breccia, spatially associated with a syenite intrusion. Fragments in the breccia zone include volcanic wall rock and syenite (J. Scott, Thunder Bay Resident Geologist's office, personal communication, 1991). Wall rocks from other zones in this area contain secondary carbonate, silica, hematite, tourmaline and pyrite. Gold mineralization also occurs in the Moss Lake area (Harris 1970), spatially associated with deformed and albitized rocks and a syenite intrusion (Chorlton 1987). In the Gold Creek area, Duckworth Township, gold occurs in quartz ladder veins which cut syenite dikes (J. Scott, Thunder Bay Resident Geologist's Office, personal communication, 1991). All these occurrences have some similarities with syenite intrusion-associated gold deposits in the Matachewan and Kirkland Lake areas.

### TECTONIC SETTING

These late to posttectonic intrusions, which range from calc-alkalic to shoshonitic, have some compositional similarities to pretectonic, calc-alkalic tonalite suite rocks (Barrie and Shirey 1991). The posttectonic intrusions were derived from enriched mantle sources (Shirey and Hanson 1984), and were emplaced into the coalesced arc and oceanic successions late during, or following, the subduction history.

## Mafic to Ultramafic Intrusions

In the Superior Province of Ontario, both tholeiitic and magnesian ultramafic to mafic, layered intrusions occur within (*see* Figure 22.12; Sutcliffe 1988): 1) granite-greenstone subprovinces; 2) plutonic subprovinces; 3) metasedimentary subprovinces; and 4) associated with subprovince boundary faults, such as along the contact between the Wabigoon and Quetico subprovinces. Two general associations are recognized (Sutcliffe and Beakhouse 1990): 1) intrusions that are comagmatic with volcanic rocks; and 2) late tectonic intrusions.

### INTRUSIONS COMAGMATIC WITH VOLCANIC ROCKS

Intrusions which are closely spatially associated with mafic and komatiitic volcanic rocks include (Sutcliffe and Beakhouse 1990): 1) concordant mafic and ultramafic sills, such as the Katimiagamak sills in the Wabigoon Subprovince (Davis and Edwards 1986); 2) layered gabbro-anorthosite complexes, such as the Bad Vermilion anorthosite in the Wabigoon Subprovince (Ashwal et al. 1983) and the Big Trout Lake intrusive complex in the Sachigo Subprovince (Whittaker 1986); and 3) komatiitic dunite-gabbro, such as the Kanichee pluton in the Abitibi Subprovince (Good 1989). Intrusions closely associated spatially with composite, intravolcanic plutonic complexes and coeval intermediate to felsic volcanic rocks (Sutcliffe and Beakhouse 1990) include: 1) layered norite to gabbronorite, such as the 2.73 billion-year-old Mulcahy gabbro in the Wabigoon Subprovince (Morrison et al. 1986); and 2) the 2.71 billion-year-old Kamiskotia Gabbroic Complex in the Abitibi Subprovince (Barrie and Shirey 1991).

Conformable peridotite-gabbro intrusions, such as the Big Trout Lake intrusive complex, commonly occur in greenstone belts and are layered and chemically fractionated. The intrusions consist of lower successions of dunite, peridotite and chromite, and upper successions of gabbro-anorthosite (Whittaker 1980, 1988a, p.35). Discordant stocks and plugs occur in plutonic subprovinces (Whittaker 1988a, p.35).

### Metallogenic Associations

Mineralization consists of: 1) chromite and nickel-copper-platinum group elements in sulphide concentrations in dunite and peridotite; and 2) disseminated nickel-copper-platinum group element sulphides in gabbroic and anorthositic pegmatites and in mafic layers in anorthosite phases (Whittaker 1988a, p.35). Sulphide zones occur near the margins of intrusions, close to contacts with country rocks or contemporaneous plutons (Sutcliffe 1988). Pegmatitic gabbroic phases, associated with many of the mineralized zones, may represent zones of enhanced magmatic volatile activity (Sutcliffe 1988). Mixing between mafic or ultramafic magmas or contamination by a granitoid component may have been important in generating pegmatitic phases and inducing sulphur saturation in the magma (Sutcliffe 1988). Intrusions with magnesium-rich cumulate sequences appear to have higher potential for nickel-copper-platinum group element resources than fractionated iron-rich intrusions (Sutcliffe 1988).

In the Sachigo Subprovince (*see* Figure 22.12), some layered ultramafic to mafic intrusions, such as the Big Trout Lake intrusive complex, and homogeneous mafic intrusions in greenstone belts, are associated with nickel-copper-platinum group element resources (Whittaker 1988a, p.35).

In the Big Trout Lake intrusive complex (Thurston et al. 1979), disseminated and massive chromite, containing platinum group elements, occurs in zones and layers in dunite and peridotite (Whittaker 1980). Laurite ($RuS_2$), and osmium- and iridium-bearing sulphide minerals occur as inclusions in chromite (Talkington et al. 1984). Higher osmium, iridium and ruthenium tenors are associated with chromite, whereas, higher platinum and palladium tenors are associated with net-textured sulphide minerals (Borthwick 1984). Other zones rich in platinum group elements include resorbed peridotite layers with associated sulphide minerals (Whittaker 1988b, p.47). Net-textured and disseminated pyrrhotite, chalcopyrite and pentlandite also occur in chromitite layers of dunite and peridotite (Whittaker 1988b, p.41). Homogeneous mafic intrusions, such as the gabbroic and diorite intrusions in the Muskrat Dam Lake greenstone belt, were emplaced into mafic volcanic successions (Ayres 1964a–i, 1965) and have undergone little fractionation (Whittaker 1988b, p.49). These intrusions are associated with few sulphide occurrences (Ayres 1964a–i, 1965).

In the Uchi Subprovince, at the Thierry Mine, in the Pickle Lake greenstone belt (*see* Figure 22.12), copper-nickel-platinum group element sulphide mineralization is localized within a shear zone in a mafic and ultramafic intrusion (Sage and Breaks 1982; Patterson and Watkinson 1984, 1988, p.57). The principal mineralization type consists of chalcopyrite, pyrrhotite, pentlandite, pyrite, violarite, merenskyite, moncheite and stützite in breccia and mylonite zones (Patterson and Watkinson 1988, p 60). Chalcopyrite-rich zones are enriched in platinum (Patterson and Watkinson 1988, p.60). In the breccia and mylonite mineralization, chalcopyrite and pyrrhotite occur in approximately equal proportions, a characteristic attributed to ductile redistribution of chalcopyrite (Patterson and Watkinson 1984).

Other fractionated gabbroic intrusions, such as the 2733 million-year-old (Morrison et al. 1985) Mulcahy gabbro in the Atikwa–Lawrence volcanic-plutonic terrane, south of Dryden, belong to an iron tholeiite suite. Tholeiitic intrusions appear to be associated with only minor occurrences of nickel-copper sulphide mineralization (Sutcliffe 1988, p.83).

In the Abitibi Subprovince, nickel-copper-platinum group element mineralization occurs in syntectonic to late tectonic, tholeiitic and komatiitic intrusions (Good 1987). The synvolcanic, komatiitic, Kanichee dunite-gabbro pluton in the Temagami greenstone belt (*see* Figure 22.12) is host to 3 habits of sulphide mineralization (James and Hawke 1984; Good 1989). Disseminated and net-textured pyrrhotite, chalcopyrite, pyrite, pentlandite occur within dunite (James and Hawke 1984). Secondary carbonate veins of redistributed magmatic chalcopyrite, pyrite and pentlandite are enriched in platinum group elements, and cut the peridotite (James and Hawke 1984; Good 1989). In the Abitibi greenstone belt, copper-nickel sulphide mineralization occurs within the 2702 ± 2 million-year-old, synvolcanic, magnesian tholeiitic, Montcalm pyroxenite-gabbro-anorthosite intrusion (Barrie and Naldrett 1990). Mineralization, consisting of disseminated, net-textured, massive, and inclusion breccia sulphide zones and very low platinum group element tenor, occurs as 2 subvertical lenses within clinopyroxene-plagioclase gabbro cumulate (Barrie and Naldrett 1990). Sulphide and oxide minerals within the mineralized zones include pyrrhotite, pentlandite and chalcopyrite accompanied by minor magnetite, pyrite and violarite (Barrie and Naldrett 1990). Igneous rocks near the mineralized zone have pegmatitic textures (Barrie and Naldrett 1990). The genesis of the sulphide mineralization is attributed to crustal contamination (Barrie and Shirey 1991) of a tholeiitic magma derived from either (Barrie and Naldrett 1990; Barrie et al. 1990): 1) partly fractionated, more primitive part of a magnesian tholeiitic magma chamber; or 2) from a chamber deeper in the crust.

## TECTONIC SETTING

Synvolcanic mafic to ultramafic intrusions likely evolved in tectonic settings similar to those of their cogenetic volcanic successions. Intrusions, such as the Montcalm Gabbroic Complex (Barrie 1990), were emplaced into stable continental settings (Barrie and Shirey 1991). Contamination by crust enriched in LREEs was an important factor which contributed to the formation of the nickel-copper sulphide mineralization (Barrie et al. 1990; Barrie and Shirey 1991). Other intrusions, such as the Kamiskotia Gabbroic Complex (Barrie 1990), evolved by low-pressure, tholeiitic fractionation from a partial melt derived from primitive mantle peridotite (Barrie and Shirey 1991). The intrusion was emplaced into a continental or extensional oceanic-arc regime, although no crustal contamination is indicated (Barrie and Shirey 1991).

The extent to which these settings are typical of large, synvolcanic mafic-ultramafic intrusions elsewhere in the Superior Province is not clear, due to the paucity of integrated, radiogenic isotopic studies. However, the tholeiitic Big Trout Lake intrusive complex and komatiitic Kanichee pluton were emplaced coevally with spatially associated metavolcanic rocks. Tectonic settings represented by this spectrum of comagmatic intrusions and volcanic rocks are complex. Rifted oceanic or back-arc settings, represented by layered mafic to tholeiitic intrusions in the Kidd–Munro or Stoughton–Roquemaure assemblages in the Abitibi greenstone belt (Jackson and Fyon, this volume) are favoured tectonic settings for some of the intrusions.

## LATE TECTONIC MAFIC-ULTRAMAFIC INTRUSIONS

A series of late tectonic, layered, ultramafic to mafic complexes occurs along the northern contact of the Quetico Subprovince (*see* Figure 22.12), within, or near, the Quetico Fault (MacTavish and Dutka 1988, p.90). Included in this array is the 2.69 billion-year-old (D.W. Davis, personal communication to R.H. Sutcliffe, reported in Sutcliffe and Beakhouse 1990) Lac des Iles intrusive complex in the Wabigoon Subprovince (Sutcliffe et al. 1989) and perhaps the Entwine Lake intrusion (Davies 1964). As well, small isolated dikes, sills and stocks of syntectonic gabbro, pyroxenite, and peridotite occur along, or near, the Quetico Fault.

Some of these intrusions were coeval with extensive tonalite plutonism (Sutcliffe 1988). Many of these intrusions exhibit features characteristic of the Mesozoic Alaskan-type ultramafic intrusions (Taylor 1967) including (MacTavish and Dutka 1988, p.91; Sutcliffe 1988, p.70): 1) an association with marginal zones of hornblende-rich ultramafic rocks; 2) variation in rock types within the complexes due to processes of multiple magma injection and some *in situ* fractional crystallization; and 3) association of ultramafic rocks with gabbroic rocks which are not cogenetic. Unlike the Alaskan-type ultramafic intrusions, these intrusions are silica saturated and contain cumulus orthopyroxene (Sutcliffe 1988, p.73). Some of these zoned intrusions are spatially and temporally associated with posttectonic intrusions of the diorite-monzodiorite suite (Sutcliffe and Beakhouse 1990).

### Metallogenic Associations

Nickel-copper-platinum group element occurrences and deposits in the Wabigoon Subprovince are principally within the Lac des Iles intrusive complex (Sutcliffe et al. 1989). This mineralization is characterized by disseminated iron-copper-nickel sulphides, containing low Pt/(Pt + Pd) ratios (Sutcliffe 1988; Sutcliffe et al. 1989). Nickel-copper-platinum group element mineralization also occurs associated with an array of smaller intrusions in the Atikokan area, along the contact between the Wabigoon and Quetico subprovinces (Sutcliffe 1988, p.71). Nickel-copper-platinum group element mineralization in these intrusions occurs as net-textured and disseminated chalcopyrite, pyrrhotite, pyrite and pentlandite in both peridotite and gabbro phases (MacTavish and Dutka 1988, p.91). Zones rich in platinum group elements are associated with peridotite (MacTavish and Dutka 1988, p.98). The mineralization generally occurs near an internal contact between intrusive rocks, or near the contact with the surrounding metasedimentary rocks (MacTavish and Dutka 1988, p.98).

Chromite occurs within a serpentinized ultramafic intrusion (*see* Figure 22.12) within a succession of metasedimentary rocks in the Quetico Subprovince at Chrome and Puddy lakes (Whittaker 1988c, p.87). Layered and disseminated chromite occurs in peridotite (Whittaker 1980). Chalcopyrite and pyrrhotite also occur with the peridotite (Simpson and Chamberlain 1967).

Within the English River Subprovince, nickel-copper resources are known in the Werner and Gordon lakes area, localized in plugs and lenses of peridotite and associated mafic rocks (*see* Figure 22.12). These intrusions were emplaced into, and along, the Werner Lake fault zone, located 20 km south of the Sydney Lake fault (Carlson 1958, p.21). Lenses, disseminations, streaks and massive sulphide bodies consist of pyrrhotite, pentlandite, chalcopyrite, violarite and pyrite (Macdonald 1988a, p.67). Cobaltite occurs in some of the sulphide mineral assemblages (Carlson 1958).

### Tectonic Setting

Mineralized intrusions localized along subprovince boundaries, such as those in the Atikokan area, occur in proximity to the Wabigoon–Quetico Subprovince boundary fault zone. These intrusions were emplaced late during the Kenoran Orogeny, along a fault zone between the Wabigoon and Quetico subprovinces (Sutcliffe 1988; Williams, Stott, Thurston et al., this volume). Collision between magmatic arcs, such as the Wabigoon Subprovince (Langford and Morin 1976) and accretionary sedimentary prism, such as the Quetico Subprovince (Percival and Williams 1989) culminated at approximately 2700 Ma (Stott et al. 1987) during the Kenoran Orogeny. Some of these intrusions show evidence of crustal contamination or mixing with coeval tonalitic magma (Sutcliffe 1989), which contributed to the genesis of associated nickel-copper-platinum group element mineralization (Sutcliffe et al. 1989). Coeval mafic and tonalitic magmatism was initiated by subduction processes, involving magma underplating (Fyon et al. 1989; Sutcliffe 1989; Sutcliffe et al. 1990). Resultant plutons were emplaced into both granite-greenstone and metasedimentary subprovinces.

The potential for these intrusions to host nickel-copper-platinum group element or chromite mineralization appears to depend more on processes such as crustal assimilation or magma mixing at the emplacement site, than the timing or tectonic setting of emplacement (Sutcliffe and Beakhouse 1990).

## "GOLD-ONLY" MINERALIZATION

"Gold-only" deposits in the Superior Province are defined as those from which gold is the principal metal extracted, generally associated with by-product silver in the ratio of 5 to 10 parts gold to 1 part silver. For convenience, these deposits are referred to as gold deposits in this section. Deposits, such as volcanic-associated, massive, base metal-bearing (VMS) deposits, where gold can be produced as a by-product metal, are not considered in this section. Many of the metallic mineral deposit types in the Superior Province of Ontario discussed so far in this chapter display spatial, temporal and possibly genetic relationships to specific intrusive rocks and/or lithostratigraphic assemblages. Such an association cannot be as easily established for gold deposits. Gold deposits do not appear in any specific rock type within a greenstone belt (Hodgson 1983). Neither has it been demonstrated conclusively that gold deposits occur more often within specific assemblage types. Many gold deposits show a spatial and temporal relationship to late tectonic elements, such as regional deformation zones (Figure 22.15), but have a less obvious temporal or genetic relationship to specific intrusion types (Colvine et. al. 1988).

Approximately 4500 t of gold have been extracted from gold deposits in the Superior Province (Table 22.7). We review the characteristics of these deposits at various scales (Tables 22.8 and 22.9). Representative deposits from Quebec are included for comparative purposes. Although the genesis of gold deposits is not discussed in detail, their genesis remains the subject of much debate (Karvinen 1978; Kerrich and Fryer 1979; Valliant et al. 1983; Phillips et al. 1984; McNeil and Kerrich 1986; Groves and Phillips 1987; Groves et al. 1987; Perring et al. 1987; Rock et al. 1989;

Colvine 1989; Gauthier et al. 1990; Marquis et al. 1990a, 1990b). Evidence is presented that is consistent with the formation of gold deposits during: 1) an orogeny at 2.86 Ga that affected the northern Sachigo Subprovince; 2) an orogeny that predates 2740 Ma in the Uchi Subprovince; 3) an orogeny at 2740 Ma in the Uchi Subprovince; and 4) during, or following, the accretionary Kenoran Orogeny (2.72 to 2.65 Ga) in the Uchi, Wabigoon, Wawa and Abitibi subprovinces. Differences in attribute characteristics between, and within, gold deposits does not necessarily mean that different genetic gold deposit types exist in the Superior Province of Ontario, although that possibility exists. These differences are interpreted to represent variations on a theme, induced by differences in physico-chemical environments at each depositional site (Colvine et al. 1988).

## Province-Scale Characteristics of Gold Camps

The term "deformation zone" is defined in the Introduction to this volume (Thurston, this volume). The Superior Province is cut by many regional deformation zones which display variable strikes (*see* Figure 22.15). These deformation zones are up to several kilometres in width and have strike lengths of hundreds of kilometres. Some deformation zones are characterized by the intense development of planar schistosity (e.g., Porcupine–Destor deformation zone, Abitibi Subprovince). Others are characterized mainly by folded rocks, accompanied by planar schistosity (e.g., Pickle Crow deformation zone, Uchi Subprovince; Table 22.10). Many gold camps display

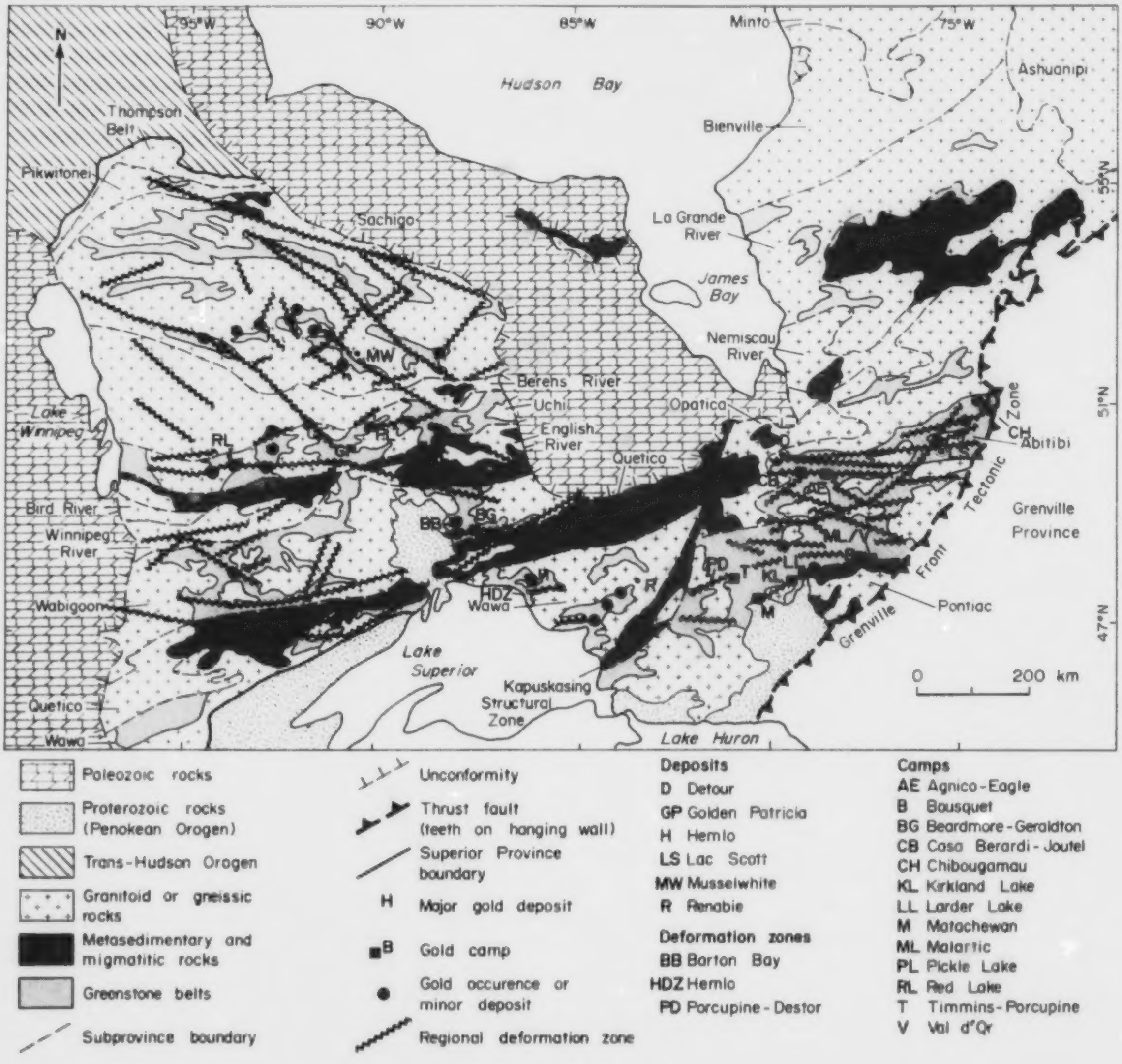
N
95°W
90°W
85°W
Minto
75°W
Hudson Bay
Ashuanipi
Thompson Belt
Bienville
Pikwitonei
55°N
Sachigo
La Grande River
James Bay
MW
Nemiscau River
Berens River
Uchi
Lake Winnipeg
English River
Opatica
RL
GP
PL
Quetico
CH
Abitibi
Zone
Tectonic
Front
Grenville
51°N
Bird River
BB
BG
CB
AE
B
V
ML
Grenville Province
Winnipeg River
PD
T
KL
LL
H
HDZ
Wawa
R
M
Pontiac
Wabigoon
47°N
Lake Superior
Kapuskasing Structural Zone
0
200 km
Quetico
Wawa
Lake Huron
Paleozoic rocks
Proterozoic rocks (Penokean Orogen)
Trans-Hudson Orogen
Granitoid or gneissic rocks
Metasedimentary and migmatitic rocks
Greenstone belts
Subprovince boundary
Unconformity
Thrust fault (teeth on hanging wall)
Superior Province boundary
H Major gold deposit
B Gold camp
Gold occurence or minor deposit
Regional deformation zone
Deposits
D Detour
GP Golden Patricia
H Hemlo
LS Lac Scott
MW Musselwhite
R Renabie
Deformation zones
BB Barton Bay
HDZ Hemlo
PD Porcupine-Destor
Camps
AE Agnico-Eagle
B Bousquet
BG Beardmore-Geraldton
CB Casa Berardi-Joutel
CH Chibougamau
KL Kirkland Lake
LL Larder Lake
M Matachewan
ML Malartic
PL Pickle Lake
RL Red Lake
T Timmins-Porcupine
V Val d'Or

**Figure 22.15.** Distribution of regional deformation zones and some spatially associated gold camps in the Superior Province.

**Table 22.7.** Approximate historic gold production plus reserves of some of the larger gold mines in the Superior Province.

| Rank | Mine (Reference)[1] | Gold Resource (Production + Reserves) | |
|---|---|---|---|
| | | Metric Tonnes (t) | Million Troy Ounces |
| 1 | Hollinger, Timmins–Porcupine (3,14) | 625 | 20 |
| 2 | Dome, Timmins–Porcupine (3) | 345 | 11 |
| 3 | McIntyre, Timmins–Porcupine (3,14) | 330 | 10.6 |
| 4 | Kerr Addison, Larder Lake (3,12) | 335 | 10.8 |
| 5 | Lake Shore, Kirkland Lake (3,12) | 265 | 8.5 |
| 6 | Williams, Hemlo (8) | 197 | 6.3 |
| 7 | Campbell Red Lake, Red Lake (1,2,15) | 318 | 10.2 |
| 8 | Golden Giant, Hemlo (8) | 180 | 5.8 |
| 9 | Wright–Hargreaves, Kirkland Lake (3,14) | 150 | 4.8 |
| 10 | Lamaque, Val d'Or (5) | 146 | 4.7 |
| 11 | A.W. White (Dickenson), Red Lake (1,2,15) | 118 | 3.8 |
| 12 | Teck–Hughes, Kirkland Lake (3,12) | 115 | 3.7 |
| 13 | Pamour No. 1, Timmins–Porcupine (4) | 107 | 3.4 |
| 14 | Sigma, Val d'Or (5) | 103 | 3.3 |
| 15 | David Bell, Hemlo (8) | 94 | 3.0 |
| 16 | Doyon, Bousquet (9) | 91 | 2.9 |
| 17 | Macassa, Kirkland Lake (5,12) | 89 | 2.8 |
| 18 | East Malartic, Malartic (7) | 88 | 2.8 |
| 19 | Madsen, Red Lake (1,2) | 75 | 2.4 |
| 20 | Bousquet (No. 1,2,3,4,5), Bousquet (9) | 69 | 2.2 |
| 21 | Aunor, Timmins–Porcupine (3,11) | 61 | 1.9 |
| 22 | Malartic Goldfields, Malartic (7) | 53 | 1.7 |
| 23 | Sylvanite, Kirkland Lake (3,12) | 53 | 1.7 |
| 24 | Hallnor, Timmins–Porcupine (3) | 50 | 1.6 |
| 25 | Camflo, Val d'Or (6) | 48 | 1.5 |
| 26 | Preston, Timmins–Porcupine (3) | 47 | 1.5 |
| 27 | Detour Lake, Detour (5) | 48 | 1.5 |
| 28 | Pickle Crow, Pickle Lake (11) | 45 | 1.4 |
| 29 | Macleod–Cockshutt, Geraldton (16) | 46 | 1.4 |
| 30 | Agnico–Eagle, Joutel (10) | 40 | 1.3 |
| 31 | Barnat–Sladen, Malartic (7) | 37 | 1.2 |
| 32 | Paymaster, Timmins–Porcupine (11) | 37 | 1.2 |
| 33 | Renabie, Missanabie–Renabie (17) | 34 | 1.1 |
| 34 | Canadian Malartic, Malartic (7) | 33 | 1.1 |
| 35 | Ross (12) | 32 | 1.0 |
| 36 | Buffalo Ankerite, Timmins–Porcupine (11) | 31 | 1.0 |
| 37 | Delnite, Timmins–Porcupine (11) | 29 | 0.95 |
| 38 | Dumagami (Main), Bousquet (9) | 25 | 0.81 |
| 39 | Central Patricia, Pickle Lake (11) | 19 | 0.63 |
| 40 | Young–Davidson, Matachewan (13) | 18 | 0.58 |
| 41 | Matachewan Consolidated, Matachewan (13) | 11 | 0.37 |
| 42 | Dona Lake, Red Lake (18) | 8.8 | 0.29 |
| 43 | McDermott, Harker–Holloway (12) | 7.7 | 0.25 |

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[1] *References: 1– Lavigne et al. 1986; 2– Andrews et al. 1986; 3– Fyon et al. 1986; 4– West et al. 1990; 5– Colvine et al. 1988; 6– Sauve and Makila 1990; 7– Sansfaçon and Hubert 1990; 8– Muir et al. 1991; 9– Marquis et al. 1990b; 10– Simard and Genest 1990; 11– Ferguson et al. 1971; 12– Bath 1990; 13– Gordon et al. 1978; 14– Wood, Burrows et al. 1986; 15– B. Atkinson, MNDM, personal communication, 1991; 16– Macdonald 1988b; 17– Sage and Heather 1991. 18– D. Janes, MNDM, personal communication, 1991.*

**Table 22.8.** Some structural characteristics of regional deformation zones and contained faults spatially associated with gold camps in the Superior Province of Ontario.

| **Camp**[1]<br>**Greenstone Belt (Reference)**[7] | **Name of Deformation Zone**[2] | **Type**[3] | **Associated Rock Assemblage**[4] | **Contained Discrete High-Strain Zone** | **Structural Characteristics of Deformation Zones**[5] | | |
|---|---|---|---|---|---|---|---|
| | | | | | **Strike** | **Dip** | **Plunge (element)** |
| Red Lake<br>Red Lake greenstone belt (1,2,22,23,26) | Cochenour–Gullrock Lake deformation zone | IGDZ | komatiite-tholeiite | — | ESE | vertical to steep to S | steep SW to down-dip (stretch lineation) |
| | Flat Lake–Howey Bay deformation zone | IGDZ | komatiite-tholeiite | — | NE to ENE | steep to SE | shallow to subhorizontal (stretch lineation)<br>subvertical stretch lineation in low-strain lozenges between high-strain zones |
| | | | | Austin shear zone | 015° to 025° | steep (070°) to SE | not well constrained |
| Pickle Crow<br>Pickle Lake greenstone belt (28, 29, 30) | Domain 1<br>Domain 2 | IGDZ | bimodal mafic-felsic volcanic rocks | — | 230° to 240° | steep (75° to 80°) NW | plunge of regional folded BIF is moderate to steep (55° to 75°) to NE |
| Hemlo Camp<br>Schreiber–Hemlo greenstone belt (14,15,19,20,31) | Hemlo deformation zone | IGDZ | greenstone belt sedimentary rocks, bimodal mafic-felsic volcanic rocks | — | belt scale: 2 orientations, E and ESE, with apparent curvilinear junction. | steep | plunge not well constrained at belt scale |
| | | | feldspathic sedimentary rocks, wackes | Cedar Creek fault | 290° | not well constrained | moderate ENE, not well constrained |
| | | | quartz-feldspar porphyry complex, feldspathic sedimentary rocks, wackes | Lake Superior shear zone | 290° | steep N (70°) | not well constrained |
| | | | Mafic iron-rich amphibolite (volcanic?), feldspathic sedimentary rocks, granitoid and mafic and ultramafic dikes. | Hemlo fault | 290° to 265° | steep to N (60° to 90°) | various; typically, moderate ENE to E crenulation |

Table 22.8. Some structural characteristics of regional deformation zones and contained faults spatially associated with gold camps in the Superior Province of Ontario.

| Camp[1]<br>Greenstone Belt (Reference)[7] | Additional Notes[8] | Examples of Associated Mines |
|---|---|---|
| Red Lake<br>Red Lake greenstone belt (1,2,22,23,26) | Heterogeneous; i) early brittle fractures filled with iron-carbonate veins and breccia zones; ii) ductile phase; iii) later brittle, Riedel shear fractures filled by iron-carbonate veins; iv) late P-shear fractures filled by iron-carbonate veins. | Campbell Red Lake, A.W. White (Dickenson) |
| | Normal down-dip to oblique slip component. | |
| | Several discrete high-strain zones, 1 to 100 m wide (e.g., Austin shear zone); at west end, zone defines an open S-shaped flexure. Sinistral horizontal slip component. | Madsen, Starratt–Olsen |
| Pickle Crow<br>Pickle Lake greenstone belt (28, 29, 30) | Domain of variable schistosity development from intense to moderate. Overturned to south isoclinal fold with associated S- and Z-shaped minor on northwest and southeast limbs respectively. | Pickle Crow and Central Patricia Mines |
| Hemlo Camp<br>Schreiber–Hemlo greenstone belt (14,15,19,20,31) | Polyphase strain history; early S-asymmetry folds; later Z-asymmetry folds; late dextral oblique component. | Williams, Golden Giant, David Bell |
| | Generally not well exposed or characterized; may be a narrow, discrete plane, but is ductile shear zone at barren sulphide zone; offsets diabase dike; last increment of slip may have been brittle. | |
| | Not well characterized. | |
| | Possible dextral reactivation of an earlier sinistral ductile high-strain zone. | |

**Table 22.8.** Some structural characteristics of regional deformation zones and contained faults spatially associated with gold camps in the Superior Province of Ontario.

| Camp[1]<br>Greenstone Belt (Reference)[7] | Name of Deformation Zone[2] | Type[3] | Associated Rock Assemblage[4] | Contained Discrete High-Strain Zone | Structural Characteristics of Deformation Zones[5] | | |
|---|---|---|---|---|---|---|---|
| | | | | | **Strike** | **Dip** | **Plunge (element)** |
| Geraldton<br>Beardmore–Geraldton greenstone belt (8,9,24,25) | Barton Bay deformation zone | SBDZ | komatiite-tholeiite; intragreenstone belt turbidite | — | 110° | steep to vertical | shallow, W-plunging fold axes, mineral and dimensional lineations |
| | | | | Bankfield–Tombill shear zone | 110° | vertical to steep S | subhorizontal slickensides, shallow W-plunging fold axes of folded foliation |
| Renabie Area<br>Eastern Michipicoten greenstone belt (10,11) | Eastern extension of the Goudreau–Lochalsh–Missanabie–Renabie composite zones of deformation | IGDZ | bimodal mafic-felsic volcanic rocks, external tonalite-trondhjemite | Renabie high-strain zone | variable, but easterly on surface, ESE at depth | steep to S | moderate to steep WSW |
| | | | | C-zone high-strain zone | 075° to 085° | moderate to steep S | moderate NW |
| | | | | Braminco No. 21 high-strain zone | 135° to 145° | moderate to steep SW | moderate W |
| Detour Area<br>Abitibi greenstone belt (7) | Detour deformation zone | IGDZ | komatiite-tholeiite, turbidite sedimentary rocks | — | W to NW (260° to 300°) | Vertical to steep N (70°) | Moderate W to NW (mineral and dimensional stretch lineation) |
| Timmins–Porcupine<br>Abitibi greenstone belt (3,4,5,6,21) | Porcupine–Destor deformation zone | IGDZ | komatiite-tholeiite, turbidite sedimentary rocks, bimodal mafic-felsic volcanic rocks, internal intrusions, alluvial-fluvial sedimentary rocks | — | WSW (245°) | steep N in Timmins camp. | Steep to down-dip (dimensional lineation) |
| | | | komatiite-tholeiite, intrusions, turbiditic sedimentary rocks, alluvial-fluvial sedimentary rocks | Dome fault | ENE (065°), locally folded in Timmins camp area | steep to S | steep to down-dip (dimensional lineation). Shallow NE-plunging chevron fold axis with NE-striking axial planar foliation |
| | | | komatiite-tholeiite, intrusions | Hollinger fault | NE to ENE (042° to 085°) | steep to S | moderate to steep NE to ENE (50° to 80°) (dimensional lineation) |
| Harker–Holloway<br>Abitibi greenstone belt (12,13) | Porcupine–Destor deformation zone | IGDZ | komatiite-tholeiite, turbidite sedimentary rocks, intrusions, alluvial-fluvial sedimentary rocks | — | E | not well constrained | not well constrained |
| | | | komatiite-tholeiite | McDermott high-strain zone | ENE (075°) | vertical to steep to S | moderate (045°) to the SW |

Table 22.8. Some structural characteristics of regional deformation zones and contained faults spatially associated with gold camps in the Superior Province of Ontario.

| **Camp**[1]<br>**Greenstone Belt (Reference)**[7] | **Additional Notes**[6] | **Examples of Associated Mines** |
|---|---|---|
| Geraldton<br>Beardmore–Geraldton greenstone belt (8,9,24,25) | Polyphase strain history, regional ductile shear with dextral, Z-asymmetrical folds. | Hard Rock, Macleod Cockshutt |
| | Ductile; dextral horizontal component of slip; locus of mafic, ultramafic and felsic dike intrusion; 100 to 300 m wide. | |
| Renabie Area<br>Eastern Michipicoten greenstone belt (10,11) | Discrete 085° high-strain zone, several metres in width. Overprinted by 110° fabric in auriferous areas. This produces S-asymmetry folds of the high-strain zone, contained quartz veins and mafic-lamprophyre dikes. Down plunge, the S-folded mineralized zone is transposed into the ESE-striking axial planar orientation. | Renabie Mine |
| | Similar to Renabie high-strain zone. Orientation of mineralized zone is ESE. | Renabie C-zone |
| | Discrete 140° high-strain zone overprinted by ESE-striking fabric that produces Z-asymmetry fold of high-strain zone, contained quartz veins, and mafic-lamprophyre dikes. | Braminco No. 21 vein |
| Detour Area<br>Abitibi greenstone belt (7) | Early ductile, followed by brittle and ductile (crenulation), and then a 2nd brittle event associated with Au mineralization. Late crenulation fold axis 170° to 190°. | Detour Lake Mine |
| Timmins–Porcupine<br>Abitibi greenstone belt (3,4,5,6,21) | Early reverse (S over N) component of slip, synchronous with early Dome fault reverse slip; reactivated during development of later NE-striking foliation. | Pamour No. 1, Hallnor |
| | Folded deformation zone; possibly early, reverse component of slip; reactivated during development of later NE-striking foliation. S-asymmetry folds. | Dome, Paymaster, Buffalo Ankerite, Aunor, Delnite |
| | Early ductile phase; later brittle fault; possibly early, reverse component of slip. | Hollinger, McIntyre, Coniaurum |
| Harker–Holloway<br>Abitibi greenstone belt (12,13) | Not well constrained in this area; 500 m wide. | Holt–McDermott, Canamax Resources Matheson Project |
| | Subsidiary deformation zone to south of Porcupine–Destor deformation zone; outer zone of ductile deformation; inner zone of brittle deformation (breccia). | |

Table 22.8. Some structural characteristics of regional deformation zones and contained faults spatially associated with gold camps in the Superior Province of Ontario.

| Camp[1]<br>Greenstone Belt (Reference)[7] | Name of Deformation Zone[2] | Type[3] | Associated Rock Assemblage[4] | Contained Discrete High - Strain Zone | Structural Characteristics of Deformation Zones[5] | | |
|---|---|---|---|---|---|---|---|
| | | | | | Strike | Dip | Plunge (element) |
| Kirkland Lake<br>Abitibi greenstone belt (16,17) | Larder–Cadillac deformation zone | IGDZ | alluvial-fluvial sedimentary rocks, komatiite-tholeiite, intrusions | — | 2 main strike orientations of 070° and 110° with curviplanar junction | generally steep to S | steeply oblique to down-dip (dimensional lineation) |
| | Kirkland Lake "main break" | IGDZ | Alluvial-fluvial sedimentary rocks, alkalic intrusions | — | ENE (060° to 080°) | Vertical to steep S, decreasing dip with depth | not well constrained |
| Larder Lake<br>Abitibi greenstone belt (18,27) | Larder–Cadillac deformation zone | IGDZ | komatiite-tholeiite, turbidite sedimentary rocks, alluvial-fluvial sedimentary rocks, intrusions | — | 2 main strike orientations of 070° and 110°, with curviplanar junction | steep S | steep SE to SW<br>In north part of zone, fold axes and dimensional lineations plunge NE to E; in south part of zone, fold axes and dimensional lineations plunge SW to W |

Table 22.8. Some structural characteristics of regional deformation zones and contained faults spatially associated with gold camps in the Superior Province of Ontario.

| Camp[1]<br>Greenstone Belt (Reference)[7] | Additional Notes[6] | Examples of Associated Mines |
|---|---|---|
| Kirkland Lake<br>Abitibi greenstone belt (16,17) | Some displacement postdates 2.675 Ga sedimentary rocks; indications of apparent horizontal dextral component slip exceed number of sinistral horizontal component of slip;south over north reverse sense of slip; part of wider zone of deformation; overprinted by ENE- to NE-striking fabric. | Baldwin Consolidated, Golden Gate in Kirkland Lake camp area. |
| | Early brittle-ductile fault with a south over north reverse component; folded (?) quartz veins; reactivated to yield late brittle component subsidiary faults and associated gouge; premineralization displacement on the main branch is rotational, vertical and amounted to approximately 400 m at the west end and 100 m at the east end with little or no horizontal displacement. | Macassa, Kirkland Lake Gold, Lake Shore, Wright–Hargreaves, Tech–Hughes, Sylvanite, Toburn |
| Larder Lake<br>Abitibi greenstone belt (18,27) | As above, generally. | Kerr Addison, Chesterville |

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[1] *A gold deposit is defined as a collection of discrete ore shoots, lenses or zones which may or may not be separated from each other by subore-grade material. The broad envelope which encloses the ore shoots, lenses, or zones defines the boundary of the deposit. A deposit may be mined from more than one mine. A gold camp is defined as one, generally large, or several, generally smaller, gold deposits that are located within a restricted geographic area. For example, the Hemlo camp consists of a single gold deposit that is exploited from 3 mines.*

[2] *Definition of deformation zone: broad, generally curviplanar zone in which there are domains characterized by intensely foliated and folded rocks. Strain within the zone is generally heterogeneous. The zone is composed of numerous discrete, narrower zones of high strain. Rocks exhibiting little deformation can occur within the deformation zone. An subjectively defined envelope, drawn to enclose the highest density of discrete high-strain zones, constitutes the deformation zone boundary.*

[3] *Type of regional deformation zone: IGDZ- intragreenstone belt deformation zone; SBDZ- subprovince boundary deformation zone.*

[4] *Rock assemblage: an informal term used to refer to a regionally mappable collection of rocks defined by lithic, structural, geochemical, geochronological, metamorphic and geophysical parameters. Contained units within an assemblage need not obey the "law of superposition" , but they generally display recognizable primary features. Assemblages may be fault-bounded. General assemblage types cited in this chapter include: komatiite-tholeiite; bimodal mafic-felsic volcanic; alluvial-fluvial sediments with or without associated alkalic volcanic rocks (e.g., Timiskaming-like); intergreenstone belt turbidite (e.g., metasedimentary subprovince); internal greenstone belt turbidite; internal granitoid; external granitoid. Where the camp is localized between within 2 or more adjacent assemblages, all types are indicated.*

[5] *Planar and linear measurements are present using the right-hand-rule convention. Plunge: 0° to 30°, shallow; 30° to 60°, moderate; greater than 60°, steep.*

[6] *Structural characteristics of the deformation zones have been simplified. More complex descriptions and chronologies are available in the cited references.*

[7] *References:1- Andrews et al. 1986; 2- Rogers, J. 1990; 3- Wood, Burrows et al. 1986; 4- Mason and Melnik 1986; 5- Kent 1990; 6- West et al. 1990; 7- Marmont 1986; 8- Macdonald 1988b; 9- Matheson and Douglas 1948; 10- Callan and Spooner 1987; 11- Heather 1989; 12- Workman 1986; 13- Bath 1990; 14- Harris 1986; 15- Kuhns et al. 1986; 16- Hawley 1948; 17- Thomson et al. 1948; 18- Smith et al. 1990; 19- Muir and Elliot 1987; 20- Hugon 1986; 21- Burrows and Spooner 1986; 22- Horwood 1940; 23- Durocher and Hugon 1983; 24- Lavigne 1983; 25- Buck and Williams 1984; 26- Durocher 1983; 27- Toogood and Hodgson 1985; 28- Ferguson 1966; 29- Stott, this volume; 30- Thomson 1939, 31- Muir et al. 1991.*

Table 22.9. Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| **Deposits**<br>**Greenstone Belt**<br>**(Reference[1])** | **Total Contained Au[2]**<br>**a) t Au**<br>**b) ore grade (g/t)** | **Mineralization Type[3]** | **Deposit[4]**<br>**(Strike, Dip, Plunge)** |
|---|---|---|---|
| Musselwhite Joint Venture<br>A) East Bay Zone | A) a) 46<br>b) 6.2 | Subvertical qtz-po-(ab)-(gt)-(calc) vlts and lenses in iron formation along 120°-striking cleavage, axial planar to folded iron formation; minor component of mineral inventory. | Folds: canoe-shaped, similar folds, hinge line has 0° to 60° NW plunge. |
| B) West Anticline | B) a) 17<br>b) 5.3 | Stratabound zones of diss sulphides in iron formation; elongate lenses, plunge parallel to folds; may contain qtz-po vns. | Axial planar cleavage to similar folds: 120° strike, vertical to steep SW dip. |
| North Caribou greenstone belt<br>(24, 28) | | | |
| A) A.W. White (Dickenson) | A) a) 118<br>b) 12.9 | Barren Fe-carb vns cut by Au-qtz vns, fractures, bxs and repl zones (A.W.White–F,SC zones; Campbell–F,A, North L zones). | The broad mineralized domain, in which these zones occur, has a SE strike, parallel to foliation, steep SE dip, and plunges moderately to steeply to the SE. |
| B) Campbell Red Lake | B) a) 318<br>b) 18.6 | Asp-qtz repl pipes and lenses. Brecciation and asp-qtz repl of Fe-carb vns and mafic host rocks, near or at folded mafic-ultramafic rock contacts. Present in greenschist- and amphibolite-facies rocks (A.W.White–ESC zone; Campbell–G,L zones). | |
| Red Lake greenstone belt<br>(1,2,15) | | Sulphide repl zones–bands, streaks and lenses of sulphide & folded qtz vns; lenses parallel to foliation (A.W.White–ESC-23 complex). | Zone has S strike, steep W dip, steep W plunge. |
| | | Fe-carb-qtz-tour-calc-sulphide-Au vns in intrusions within or adjacent to high-strain zones. | |
| Madsen, Red Lake<br>Red Lake greenstone belt<br>(1, 22, 23, 32) | a) 75<br>b) 9.1 | Broad mineralized domain in which several *en échelon*, diss sulphide-bearing zones of sheared rock associated with qtz vlts occur that lie at a 10° to 20° angle to the strike of the boundaries to the host high-strain zone (Austin shear zone). | Mineralized domain strikes NNE (015° to 020°), dips steeply to the SE (70°), and plunges moderately (45° to 50°) to the E. Mineralized domain is approximately parallel to the strike and dip of the host high-strain zone (Austin shear zone). |

Table 22.9. Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| Deposits<br>Greenstone Belt<br>(Reference[1]) | Ore Shoots<br>(Strike, Trend, Plunge) | Metallic Ore Minerals[5] | Host Rock | Alteration Types or Minerals Associated with Mineralization[5] | Metamorphic Grade and Alteration Mineralogy of the Associated High-Strain Zone[5] |
|---|---|---|---|---|---|
| Musselwhite Joint Venture<br>A) East Bay Zone | Best developed in areas of smaller folds within larger antiform; parallel to axial planar cleavage: 120° strike, steep SW dip. | po, (asp), (gn), (sp), (shee), (py), (cp), (Au), (alt) | qtz-mt iron formation and gt-bi-grun pelite (local hb-rich, grun-free layers); | grun, bio, stlp | amphibolite |
| B) West Anticline<br>North Caribou greenstone belt (24, 28) | Generally elongate parallel to 120° cleavage and parallel to iron formation compositional mesobands; lenses plunge parallel to fold axes (0° to 60° NW). | po, (asp), (Au) | Au-bearing ab porphyry dikes | grun, calc, ab, gt, sil | |
| A) A.W. White (Dickenson) | SE strike parallel to foliation, moderate to steep SW dip, steep SW plunge. | Au, asp | greenschist metamorphic-grade tholeiites and qtz-gabbro | sil, ser, bio, mus, Fe-carb | amphibolite; basalt–act, bio, mus, cltd, ay, gt, staur, cord, and |
| B) Campbell Red Lake | SE strike parallel to foliation, steep SW dip, steep SW plunge. | asp, Au, (py), (po) | greenschist- and amphibolite-facies tholeiites | | |
| Red Lake greenstone belt (1,2,15) | Lenses within zone have SSE strike and a steep SW to W plunge. | py, po, Au | upper greenschist- to lower amphibolite-facies tholeiites | | |
| | | Au | intrusive rocks within or adjacent to high-strain zone | | |
| Madsen, Red Lake<br>Red Lake greenstone belt (1, 22, 23, 32) | Discrete ore lenses have a NE strike (030° to 035°), steep (75°) SE to vertical dip, and down-dip to steep (78°) E plunge. The ore lenses lie at 10° to 20° angle to the envelope of the host high-strain zone and plunge steeply within the mineralized domain. | py, po, asp, sp, cp, mt, Au, (shee), (mo) | tholeiite, ultramafic rock, felsic dikes, intermediate to felsic tuff | variable; sil, chl, carb, bio, gt | metamorphic grade decreases to NE, along deformation zone<br>amphibolite to SW: basalt– and, staur, gt, calc, cltd, bio, qtz.<br>greenschist to NE: basalt– ser, bio, qtz, Fe-carb |

Table 22.9. Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| **Deposits**<br>**Greenstone Belt**<br>**(Reference[1])** | **Total Contained Au[2]**<br>**a) t Au**<br>**b) ore grade (g/t)** | **Mineralization Type[3]** | **Deposit[4]**<br>**(Strike, Dip, Plunge)** |
|---|---|---|---|
| A) Pickle Crow<br>B) Central Patricia<br>Pickle Lake greenstone belt (25,26) | A) a) 45<br>b) 15.3<br>B) a) 19<br>b) 12.5 | 2 generations of qtz vns.<br>1) Banded qtz-Fe-carb-ab-tour-shee-py(po) vns; bands consist of altered wall rock and are oriented parallel to wall rock; qtz crystals adjacent to wall rock orthogonal to wall; some textures consistent with crack-seal origin; veins are folded, locally cataclastized on a grain-scale and strike at an angle to bedding and foliation; only parts of vns carry economic concentrations of Au. | Mineralization spatially associated with nose or limbs of regional syncline defined by folded iron formation. Vns on south limb of major syncline are Z-folded. Vns on north limb of syncline are S-folded. Most vns strike 060° to 085° (15° to 40° clockwise from axial planar foliation to major syncline). |
| | | 2) Ladder qtz-calc vns that cut across at a high angle to contact with wall rock, but terminate at wall rock contact. | Most mineralization occurs within the qtz vns; very little occurs within basalt wall rock. |

Table 22.9. Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| Deposits Greenstone Belt (Reference[1]) | Ore Shoots (Strike, Trend, Plunge) | Metallic Ore Minerals[5] | Host Rock | Alteration Types or Minerals Associated with Mineralization[5] | Metamorphic Grade and Alteration Mineralogy of the Associated High-Strain Zone[5] |
|---|---|---|---|---|---|
| A) Pickle Crow<br>B) Central Patricia<br>Pickle Lake greenstone belt (25,26) | Ore shoots spatially associated with zones of folded qtz vn containing cataclastized qtz; barren parts of the vns are not fractured or cataclastized. Ore shoots have short horizontal strike, but have persistent down-plunge extent (4 to 15 times horizontal length).<br>Central Patricia (No.2 operation) Mine<br>McCallum No. 6 vn (750 foot level subsurface)<br>strike: 080°<br>dip: steep<br>plunge: steep (70°) NNE<br>Springer No. 6 vn (surface)<br>strike: 255° to 300°<br>dip: steep (65°)<br>plunge: steep NNE<br>Pickle Crow Mine<br>No. 1 (Howell vn) (1800 foot level subsurface)<br>strike: 260°<br>dip: steep (75°) N<br>plunge: steep (70°) NNE (025°) (same plunge of ore shoot and folded parts of vn)<br>No. 2 vn (750 foot level subsurface)<br>strike: variable (folded) 260°<br>dip: moderate (45°) NNW<br>plunge: 40° NE<br>No. 5 vn (1800 foot level subsurface)<br>strike: 070°<br>dip: steep<br>plunge: steep (70°) NNE<br>No. 6 vn (1800 foot level subsurface)<br>strike: 080°<br>dip: folded: moderate (50°) to NNW and SSE<br>plunge: steep (065°) NNE<br>No. 9 vn (1800 foot level subsurface)<br>strike: 125°<br>dip: steep (80°)<br>plunge: steep (75°) NNW | 2 generations of py (fractured early py carries Au, later py locally which does not carry Au). Asp abundant where wall rock is BIF. Po abundant in BIF where cut by qtz vn. Shee locally. (cpy), (gn) (sp). Gn spatially associated with Au | basalt, BIF, porphyry | basalt: chl<br>BIF: po<br>porphyry: ser, Fe-carb, tour | greenschist |

Table 22.9. Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| **Deposits Greenstone Belt (Reference[1])** | **Total Contained Au[2]**<br>**a) t Au**<br>**b) ore grade (g/t)** | **Mineralization Type[3]** | **Deposit[4] (Strike, Dip, Plunge)** |
|---|---|---|---|
| A) Pickle Crow } Continued<br>B) Central Patricia<br>Pickle Lake greenstone belt (25,26) | | Qtz-sulphide zones in folded and fractured BIF on Central Patricia near the nose of a major syncline; saddle-shaped zone developed where qtz vns cut BIF; Au in vns and sulphide BIF. | NNW strike, but saddle shaped; bisecting plane to dip dips about 75° N;<br>plunge: steep (65°) ENE. |
| Dona Lake<br>Pickle Lake greenstone belt (27, 29) | a) 5.4<br>b) 9.02 | Replacement mineralization occurs within folded BIF; qtz vns virtually absent. | 355° strike; 60° dip; 60° down-dip plunge. |
| Hard Rock–Macleod Cockshutt<br>Beardmore–Geraldton greenstone belt (7, 8) | a) 51<br>b) 4.3 | North zone: diss, repl sulphide, associated with minor component of qtz vlts, in Z-folded oxide-facies iron formation.<br>South zone: qtz-tour vns and vn envelope py in fractured ab porphyry dike. | Mineralization related to Z-folded, W-plunging units; less mineralization down-plunge where fold becomes more open. |
| Hemlo Deposit (Williams, Golden Giant, David Bell mines)<br>Schreiber–Hemlo greenstone belt (14, 16) | a) >616<br>b) 7.7 | Tabular zones of diss py associated with some highly deformed qtz vns; western part splits into several small, low-grade zones; eastern part narrows, but maintains grade. | Deposit has WNW strike, moderate to steep NNE dip, moderate NW plunge.<br>To the west, deposit has a W strike, and a moderate to steep N dip. |
| Renabie<br>Michipicoten greenstone belt (9,10,11) | a) 34.7<br>b) 6.1 | Deformed qtz vns, diss py in vn envelope.<br>Mineralization related to S-folded E-striking high-strain zone having ESE-striking axial planar foliation; down-plunge, folded high-strain zone is transposed into ESE foliation. The ore zone is narrow in upper levels, but thicker in lower levels, in the ESE plane. | E to ESE strike (060° to 100°), moderate to steep S dip, steep W to WNW plunge. |
| Hollinger–McIntyre<br>Abitibi greenstone belt (3,4) | a) 995<br>b) 9.9 | Qtz-Fe carb vns, vn envelope and diss py | ENE strike (060°), vertical to steep S dip, moderate E to ENE plunge (55°). |

Table 22.9. Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| Deposits<br>Greenstone Belt<br>(Reference[1]) | Ore Shoots (Strike, Trend, Plunge) | Metallic Ore Minerals[5] | Host Rock | Alteration Types or Minerals Associated with Mineralization[5] | Metamorphic Grade and Alteration Mineralogy of the Associated High-Strain Zone[5] |
|---|---|---|---|---|---|
| A) Pickle Crow } Continued<br>B) Central Patricia<br>Pickle Lake greenstone belt (25,26) | Detail unavailable. | po, asp, py, Au | oxide-facies BIF | po after mt | greenschist |
| Dona Lake<br>Pickle Lake greenstone belt (27, 29) | Detail unavailable. | po after mt, Au, (cpy), (sp) | oxide-facies BIF | po after mt | amphibolite: gt, grun, calc |
| Hard Rock–Macleod Cockshutt | North zone: W strike (265°), dip uncertain (steep to N?), shallow to moderate (15° to 37°) W plunge. | North zone: Au, py, asp, (po). | oxide-facies iron formation, wacke, mafic volcanic rocks | sulph,<br>sil,<br>Fe-carb | greenschist |
| Beardmore–Geraldton greenstone belt (7, 8) | South zone: W strike (270°), dip uncertain (steep to N?), shallow (16°) W plunge. | South zone: Au, py, shee, po, asp, cp | ab porphyry | sil, carb | |
| Hemlo Deposit (Williams, Golden Giant, David Bell mines)<br>Schreiber–Hemlo greenstone belt (14, 16) | Not well constrained, but possible WNW strike, moderate to steep NNE dip, steep plunge. | diss py, Au, mo, stib, real, ci, and others | felsic volcanic rocks, sedimentary rocks, subvolcanic qtz-fel porphyry (QFP) intrusion. Cut by mafic, intermediate, and felsic dikes, but dikes are not ore-grade. | mrc, ser, sil, bio, carb, ab, ky, sill | amphibolite: basalt– hb, plag, bio<br>sediments– qtz, bio, fel, gt, ser, ky, (staur), (cltd) QFP– qtz, plag, mrc, ser, bio. |
| Renabie<br>Michipicoten greenstone belt (9,10,11) | Ore shoots have same attitudes as deposit. | py, Au, tell, cp, gn | trondhjemite, tonalite, lamprophyre, mafic dikes | Qtz, ser, hem, carb, chl, ab, py, anh | greenschist: ser, chl |
| Hollinger–McIntyre<br>Abitibi greenstone belt (3,4) | Genally steep dip (75° to vertical); trend on plunge of shoots not detailed. | py, Au, shee, (cp), (sp), (gn), (po), (asp) | Fe- and Mg-tholeiitic basalt, carbonaceous sediments, qtz-fel porphyry intrusion, albitized lamprophyres | Fe-carb, ser, ab, py | greenschist: basalt– chl, act, ep, ab, calc |

**Table 22.9.** Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| **Deposits**<br>**Greenstone Belt**<br>**(Reference[1])** | **Total Contained Au[2]**<br>**a) t Au**<br>**b) ore grade (g/t)** | **Mineralization Type[3]** | **Deposit[4]**<br>**(Strike, Dip, Plunge)** |
|---|---|---|---|
| Pamour No. 1<br>Abitibi greenstone belt<br>(5, 30) | a) 107<br>b) 3.46 | Qtz-Fe-carb vns, vn envelope py. | ENE strike (070°), steep dip, moderate E to ENE plunge (60°). |
| McDermott<br>Abitibi greenstone belt<br>(12,13) | a) 7.7<br>b) 4.9 | Lobate pods (in the plunge direction) of diss py within sil and Fe-carb rock; brittle fracture-fill. | WSW strike (255°), steep to vertical dip, shallow (45°) plunge to W. |
| Macassa<br>Abitibi greenstone belt<br>(17,18,20,21,31) | a) 89<br>b) 13.7 | Qtz vns, sheeted vlts, qtz stockwork zones, tension fractures, and qtz vn bxs localized within and adjacent to Kirkland Lake "main break" and associated subsidiary faults; generally strike parallel to "main break"; early vn qtz deformed, recrystallized, and re-fractured and filled with mo-graphite, py-(cp), and qtz-ank-sulphate; followed by Au, tell; late qtz-calc-hem vlts. | WSW strike (250°), shallow S dip (35° to 45°) in upper levels, steep dip in lower levels, steep westerly plunge. |
| Kerr Addison Chesterville<br>Abitibi greenstone belt<br>(19) | a) 335<br>b) 9.3 | Green carb ore: qtz-ank-dol-carb-(Ab) vns; downward-tapered funnel-shaped alteration envelope encloses ore shoots. | WSW (240°) strike, steep (80°) N dip, steep (70°) NNE plunge. |
| | | Albitite ore. | ENE (060°) strike, steep (80°) N dip, steep (70°) E plunge. |
| | | Flow ore: diss py and py microveining and associated qtz-ank-(ab) vns with py-carb-qtz repl halo. | ENE (060°) strike, steep (80°) N dip, steep (70°) E plunge. |
| | | Graphitic ore: py-graphite-filled fractures in faulted graphitic sediments. | Similar to flow-ore. |

Table 22.9. Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| Deposits<br>Greenstone Belt<br>(Reference[1]) | Ore Shoots<br>(Strike, Trend, Plunge) | Metallic Ore Minerals[5] | Host Rock | Alteration Types or Minerals Associated with Mineralization[5] | Metamorphic Grade and Alteration Mineralogy of the Associated High-Strain Zone[5] |
|---|---|---|---|---|---|
| Pamour No. 1<br>Abitibi greenstone belt (5,30) | a) ENE trend, 35° plunge for shoots in conglomerate and wacke.<br>b) WSW trend; 45° to 50° plunge for shoots in volcanic rocks(?). | py, po, asp, (cp), (sp) | Mg-tholeiitic basalt, komatiite, wacke, arenite, conglomerate | Fe-carb, ser, qtz, py | greenschist: basalt–chl, act, ep, ab, calc |
| McDermott<br>Abitibi greenstone belt (12, 13) | | py, Au | tholeiitic basalt | sil, Fe-carb, py, hem, ab | greenschist |
| Macassa<br>Abitibi greenstone belt (17, 18, 20, 21, 31) | "main break" ore<br>surface to 3000 foot level: 250° trend; 35° to 40° plunge.<br>3000 foot to 4375 foot level: 250° trend; steep plunge.<br>4375 foot to 5300 foot level: 070° trend; "slight" east plunge. | Au, alt, cal, col, (mel), (pet), py, cp, mo | augite syenite, (felsic) syenite, fel porphyry, and qtz-fel porphyry, conglomerate, wacke, alkalic tuffs | 1) early ti after mt, hem after mt, minor bio related to cooling intrusions; pre Au<br>2) green phlogopite, oxides, sulphates, K-fel prevalent near "main break"; pre Au<br>3) fracture-vn related K-fel, ser, Fe-carb, qtz, py after hem and mt close to vns | greenschist |
| Kerr Addison Chesterville<br>Abitibi greenstone belt (19) | ENE trend (060°), steep plunge to NNE. | Au, (py), (gn), (cp), (asp), (ml) | komatiite | Komatiite: early Fe-carb-talc-chl; later Au-related fuchsite-Fe-carb-qtz | greenschist |
| | ENE trend (060°), steep plunge to NNE. | similar to green carb ore | albitized mafic dikes (lamprophyres?) | | |
| | a) ENE trend (060°), steep (70°) NNE plunge.<br>b) ENE trend 060°), moderate (45°) W plunge. | py, (cp), (sp), (asp), (gn) | Fe-tholeiitic basalt | basalt: sil, Fe-carb, py-qtz-carb vn envelopes | |
| | Similar to flow-ore. | py | carbonaceous sediments | | |

Table 22.9. Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| **Deposits**<br>**Greenstone Belt**<br>**(Reference$^1$)** | **Total Contained Au$^2$**<br>**a) t Au**<br>**b) ore grade (g/t)** | **Mineralization Type$^3$** | **Deposit$^4$**<br>**(Strike, Dip, Plunge)** |
|---|---|---|---|
| Detour Mine<br>Abitibi greenstone belt (6) | a) 48<br>b) 4.4 | Early barren qtz-calc-ep-chl vns; strongly folded. | 080° strike, vertical to steep N dip (70°), moderate W plunge (40° to 50°). |
| | | Main auriferous qtz-sulphide-fel-carb-act vns; typically sheeted; multiple fracturing and filling episodes; local wall rock bx. | |
| | | Late pink K-fel-plag-calc-dol-qtz-preh-(bi)-(chl) vns; rare sulph; local bx of wall rock; occupy late tension fractures. | |

**Table 22.9.** Generalized characteristics of some Archean gold deposits in the Superior Province of Ontario.

| Deposits<br>Greenstone Belt<br>(Reference[1]) | Ore Shoots<br>(Strike, Trend, Plunge) | Metallic Ore Minerals[5] | Host Rock | Alteration Types or Minerals Associated with Mineralization[5] | Metamorphic Grade and Alteration Mineralogy of the Associated High-Strain Zone[5] |
|---|---|---|---|---|---|
| Detour Mine<br>Abitibi greenstone belt (6) | 040° to 060° strike. | no sulphides | Fe-tholeiite, ultramafic rocks, turbidites, qtz-fel trondhjemite and diorite dikes | | Outside high-strain zone: amphibolite (hb, plag, calc, ep, qtz, mt, ti) |
| | Foliation parallel and oblique; boudinaged and folded into S- and Z-shapes.<br>Details of plunge not available. | po, py, cp, (sp), (pent) | | basalt:<br>a) pink sil halo: qtz, K-fel, ep, plag, chl, ser, preh<br>b) green ser halo: qtz, ser, (ep), (bio), (chl)<br>c) bio halo: qtz, plag, K-fel, bio-phlogopite, (ep), (ser) | Within high-strain zone: hb ep, act, chl, bio, talc, ser, mt, qtz, gt, ti, calc, dol |
| | N strike | po, py, cp, Au, (sp) | | qtz, carb, ti, preh | |

*Notes:*

*[1] References: 1- Andrews et al. 1986; 2- Rogers 1990; 3- Wood, Burrows et al. 1986; 4- Mason and Melnik 1986; 5- West et al. 1990; 6- Marmont 1986; 7- Macdonald 1988b; 8- Matheson and Douglas 1948; 9- Callan and Spooner 1987; 10- Heather 1989; 11- Sage and Heather 1991; 12- Workman 1986; 13- Bath 1990; 14- Harris 1986; 15- B. Atkinson, MNDM, 1991, personal communication; 16- Kuhns et al. 1986; 17- Hawley 1948; 18- Thomson et al. 1948; 19- Smith et al. 1990; 20- Meyer et al. 1990; 21- Hattori and Hodgson 1990; 22- Durocher 1983; 23- Horwood 1940; 24- Hall and Rigg 1986; 25- Ferguson 1966; 26- Stott and Corfu, this volume; 27- Cohoon 1986; 28- Northern Miner, 1990, November 15, p. B-16; 29- D. Janes, MNDM, personal communication, 1991; 30- P. Olson, personal communication, 1991; 31- Charlewood 1964; 32- Mathieson and Douglas 1984.*

*[2] Total contained gold data include past production and present proven and probable reserves, where available.*

*[3] Mineral habits: bx- breccia; diss- disseminated; repl- replacement; vlts- veinlets; vn- vein; vns- veins.*

*[4] The deposit consists of all discrete mineralized ore shoots or lenses. The envelope that encloses all these discrete ore shoots or zones defines the deposit boundary. Regions lacking economic mineralization may occur between discrete ore shoots, within the boundaries of the deposit.*

*[5] Mineral Abbreviations: ab- albite; act- actinolite; alt- altaite; and- andalusite; anh- anhydrite; ank- ankerite; ay- anthophyllite; asp- arsenopyrite; Au- gold; BIF- banded iron formation; bio- biotite; bn- bornite; cal- calaverite; carb- carbonate; chl- chlorite; cltd- chloritoid; ci- cinnabar; col- coloradoite; cord- cordierite; cp- chalcopyrite; calc- calcite; dol- dolomite; ep- epidote; Fe carb- ferroan carbonate; fel- feldspar; gn- galena; grun- grunerite; gt- garnet; hb- hornblende; hem- hematite; ky- kyanite; mrc- microcline; mel- melonite; ml- millerite; mo- molybdenite; mt- magnetite; mus- muscovite; pent- pentlandite; pet- petzite; plag- plagioclase; po- pyrrhotite; preh- prehnite; py- pyrite; qtz- quartz; QFP- quartz-feldspar porphyry; real- realgar; shee- scheelite; ser- sericite; sil- silicification; sill- sillimanite; sp- sphalerite; staur- staurolite; stib- stibnite; stlp- stilpnomelane; sulph- sulphidation; tell- telluride; ti- titanite; tour- tourmaline. Mineral in ( ) indicates presence, but in subordinate amount.*

a spatial association with some segments of these regional deformation zones. A few gold camps (e.g., Central Patricia and deposits in the western Lake of the Woods area) appear to be localized in, or adjacent to, contact strain aureoles adjacent to intrusions.

### ASSOCIATION WITH REGIONAL DEFORMATION ZONES AND GREENSTONE BELTS

A multitude of regional deformation zones occur in the Superior Province, and many Archean gold camps are spatially associated with some of these zones (*see* Figure 22.15, Table 22.8). Although the gold-associated regional deformation zones extend for hundreds of kilometres and cut sedimentary, plutonic and volcanic rock types, virtually all camps occur in, or immediately adjacent to, greenstone belts within the granite-greenstone subprovinces (*see* Figure 22.15; Colvine et al. 1988; Osmani and Stott 1988; Card et al. 1989; Poulsen and Robert 1989). Only some deposits in the Malartic camp, Quebec, occur outside of a granite-greenstone subprovince, within the metasedimentary Pontiac Subprovince (Sansfaçon and Hubert 1990). However, even these deposits occur immediately adjacent to the Abitibi greenstone belt, within a regional deformation zone that affects the greenstone belt rocks. Gold-associated segments of the regional deformation zones display 2 strike orientations when viewed at a greenstone belt scale: 1) northeast to east-northeast (e.g., Timmins–Porcupine, Kirkland Lake, Harker–Holloway, Madsen–Starratt–Olsen–Howey–Hasaga, and Pickle Lake camps; Hodgson and Troop 1988); and 2) east-southeast to southeast (e.g., Detour, Hemlo, Missanabie–Renabie, Geraldton camps, and the Campbell Red Lake–Dickenson deposits in the Red Lake camp).

### SUBPROVINCE BOUNDARY VERSUS INTRAGREENSTONE BELT DEFORMATION ZONES

Most gold camps occur along intragreenstone belt deformation zones (*see* Figure 22.15, Tables 22.8 and 22.9), such as the Red Lake, Hemlo, Timmins–Porcupine, Kirkland Lake and Larder Lake camps in Ontario, and the Casa Berardi–Joutel and Chibougamau camps in Quebec. These intragreenstone belt deformation zones commonly separate greenstone domains of disparate assemblage types and ages (e.g., Porcupine–Destor deformation zone: Hubert et al. 1984; Jackson and Fyon, this volume) and may represent splays of late-stage, large-scale, subprovince boundary deformation zones (Williams, Stott, Thurston et al., this volume). Some of these regional intragreenstone belt deformation zones, such as the Missanabie–Renabie deformation zone, pass from the greenstone belts into adjacent external granitoid domains.

Other gold-associated regional deformation zones correspond with, or are adjacent to, subprovince boundary faults (*see* Figure 22.15 and Table 22.8). One such deformation zone, the Barton Bay Deformation Zone, separates the Quetico and Wabigoon subprovinces (Williams 1986, 1987; Devaney and Williams 1989). The Rouyn–Val d'Or Malartic segment of the Larder–Cadillac deformation zone separates the Abitibi and Pontiac subprovinces in Quebec (Byers and Gill 1948; Robert 1990; Gauthier et al. 1990; Marquis et al. 1991a, 1991b). These regional deformation zones were active late in the tectonic history of a greenstone belt or subprovince, and record both vertical and horizonal components of slip (Robert 1989; Marquis et al. 1990a, 1990b; Williams, Stott, Thurston et al., this volume).

## Camp- and Deposit-Scale Characteristics

Gold deposits generally consist of several mineralized zones that can be mined as individual ore shoots or stopes. The boundaries of the shoots are often defined by economic considerations. Consequently, mineralization, although of subeconomic grade, is generally more areally extensive than the discrete ore shoot, but because of lower tenor, not all mineralized rock can be mined. Volumes of rock virtually devoid of mineralization typically separate ore shoots. The envelope that encloses these ore shoots and less well mineralized rock define the boundaries of a deposit. Camps consist of one or more deposits within a restricted geographic area. The Hemlo camp consists of a single deposit, mined at 3 mines, whereas the Timmins–Porcupine camp consists of several deposits, mined at approximately 25 mines.

### STRUCTURAL SETTING

Adjacent to the regional deformation zones, subsidiary sets of brittle-ductile high-strain zones commonly occur (Robert 1990), historically known as splay-structures in the Abitibi greenstone belt. The subsidiary high-strain zones may be parallel to, or oriented at an angle to, and may merge asymptotically into, the regional deformation zone. Like the regional deformation zones, the subsidiary high-strain zones were active late in the tectonic history of a greenstone belt, and record both vertical and horizontal components of slip (Robert 1989; Marquis et al. 1990a, 1990b; Williams, Stott, Thurston et al., this volume).

While the subsidiary high-strain zones have been interpreted to represent splay-structures off regional deformation zones, few systematic and rigorous studies have addressed this relationship. East-northeast-striking splay-structures in the Kirkland Lake–Larder Lake area have different kinematic characteristics from, overprint, and offset, the east-striking Larder Lake–Cadillac deformation zone (Hodgson and Hamilton 1988; Hodgson et al. 1990). These features illustrate that along the Larder Lake–Cadillac deformation zone, the structural and temporal relationships between the regional deformation zones and the spatially associated subsidiary splay-structures are complex. Thus, to interpret the subsidiary high-strain zones as splay structures related to the regional deformation zones, by virtue of a spatial association, may be simplistic.

Major gold deposits (those containing more than 30 t of gold) occur within 10 km of regional deformation zones, but do not occur along the entire lengths of the zones (Hodgson 1983; Colvine et al. 1988; *see* Figure 22.15 and Tables 22.8 and 22.9). Smaller gold deposits and occurrences (those containing less than 30 t of gold) can occur both proximal to (less than 10 km), and distal to (more than 10 km), the regional deformation zones, but are always associated with a subsidiary high-strain zone in, or immediately adjacent to, a greenstone belt (*see* Figure 22.15 and Tables 22.8 and 22.9). Some deposits, such as the Campbell Red Lake and A.W. White in the Red Lake camp (Cochenour–Gullrock Lake and Flat Lake–Howey Bay deformation zones) occur along specific segments of the regional deformation zones. Other deposits, such as the Hollinger–McIntyre–Coniaurum in the Timmins–Porcupine camp or McDermott in the Harker–Holloway camp (*see* Tables 22.8 and 22.9) occur within subsidiary high-strain zones, adjacent to the regional deformation zones.

### ASSOCIATED ASSEMBLAGES

Within greenstone belts, gold camps are associated with several different assemblage types, including (*see* Table 22.8): 1) komatiite-tholeiite volcanic (e.g., Timmins–Porcupine); 2) bimodal mafic-felsic volcanic (e.g., Missanabie–Renabie); 3) internal greenstone belt turbidite (e.g., Hemlo, Timmins–Porcupine); 4) alluvial-fluvial sedimentary (e.g., Kirkland Lake); 5) internal granitoid stock (e.g., Camflo deposit in Malartic camp); 6) quartz-feldspar porphyritic dike or stock; and 7) external granitoid intrusion (e.g., Renabie deposit in Missanabie–Renabie camp). Some, such as the Timmins–Porcupine and Geraldton camps, are localized within more than one volcanic or sedimentary assemblage (*see* Table 22.8).

### HOST ROCK ALTERATION

A variety of alteration minerals are associated with gold mineralization in the various camps (*see* Tables 22.8 and 22.9). Alteration types recognized include carbonatization, potassium metasomatism, silicification, hematitization and sulphidation. This variability is attributed, in part, to the reaction of the hydrothermal fluid with diverse rock types having different bulk compositions (Colvine et al. 1988). In addition, ambient regional metamorphic grade can vary across a camp, resulting in a corresponding difference in hydrothermal mineral assemblages that formed prior to, or synchronous with, peak metamorphic thermal and pressure conditions.

Mineral assemblages observed in camp-scale alteration zones, both within, and outside of, the regional deformation zones that are not spatially associated with gold mineralization, generally reflect the ambient metamorphic grade. Assemblages specific to the various bulk compositions and metamorphic conditions of the protoliths are listed in Table 22.9. Hematitization is characteristic of altered intrusions, specifically, but not exclusively, those having alkalic affinity (e.g., Kirkland Lake camp, Abitibi greenstone belt). Amphibolite-facies mafic and felsic rocks within the regional deformation zones can contain aluminous minerals, (e.g., Hemlo, Detour and part of Red Lake camp; *see* Table 22.9). Sulphidation is characteristic of altered oxide-facies iron formation, regardless of whether the alteration took place under amphibolite or greenschist metamorphic conditions.

Pressure and temperature conditions inferred from the hydrothermal mineral assemblages within regional deformation zones not spatially associated with gold mineralization, are generally similar to those for metamorphosed rocks outside of the deformation zone (e.g., Red Lake camp; Andrews et al. 1986). This implies that a component of deformation and hydrothermal alteration predated, or was contemporaneous with, peak metamorphism (Andrews et al. 1986; Marmont 1986; Marquis et al. 1990a, 1990b). Some of this hydrothermal alteration within the regional deformation zones that is not spatially associated with the gold mineralization, predated gold mineralization (Marmont 1986; Marquis et al. 1990a, 1990b).

## Deposit- and Stope-Scale Characteristics

In general, gold deposits in the Superior Province are characterized by complex structural styles, alteration assemblages and mineralization sequences. Differences in these features are related, largely, to the deformation history and host rock type. During deformation, each rock type responds differently; dependent, in part, on its competency relative to different adjacent rock types due to differences in bulk composition and mineralogy. Hydrothermal alteration can modify the bulk composition and resultant mineralogy of a rock and, hence, can affect its inherent competency.

### HOST ROCKS

All rock types within a greenstone belt can host gold mineralization where deformed (*see* Table 22.9; Hodgson 1983; Colvine et al. 1988). Although metavolcanic units are common host rocks (*see* Table 22.9), mineralization also occurs in chemical and clastic metasedimentary rocks (*see* Table 22.9). Intrusive rocks commonly occur near, and are mineralized in, many gold deposits. Mineralized intrusive rocks include: 1) older (*circa* 3.0 Ga; Davis and Jackson 1988) tonalite gneissic intrusions (e.g., Marmion Lake batholith; Wilkinson 1982; Stone and Kamineni 1989; Stone et al., in press) that may have been synvolcanic to older volcanic rocks; 2) younger (*circa* 2.7 Ga) synvolcanic trondhjemite-granodiorite (e.g., Mooshla and Doyon Mines, Bousquet camp, Quebec; Savoie et al. 1986); 3) synvolcanic diorite-quartz-diorite (e.g., Sigma Mine, Val d'Or camp, Quebec; Robert et al. 1983); 4) external tonalite-trondhjemite intrusions (e.g., Renabie Mine, Missanabie–Renabie camp; Heather 1989); 5) postvolcanic quartz- and feldspar-porphyritic granodiorite (e.g., Hollinger–McIntyre–Coniaurum deposit, Timmins–Porcupine camp—Wood, Burrows et al. 1986); 6) late monzonite or syenite that were emplaced following most of calc-alkalic volcanism (e.g., Canadian Arrow—Cherry 1983; Kirkland Lake camp—Burrows and Hopkins 1925, Charlewood 1964); and 7) late lamprophyre dikes (e.g., Hollinger–McIntyre–Coniaurum deposit, Timmins–Porcupine camp—Wood, Burrows et al.

1986; Kerr Addison–Chesterville deposit, Larder Lake camp—Smith et al. 1990). External granitoid intrusions, immediately adjacent to the greenstone belts, appear to be disproportionately unmineralized, although the Renabie deposit (Sage and Heather 1991) occurs in such an intrusion. Although occurrences are found in the Marmion Lake batholith, few of these properties have yielded more than several thousand tonnes of mineralized rock (Wilkinson 1982).

## ROCK ALTERATION

Rock alteration associated with gold deposits in the Superior Province is characterized by the addition of $CO_2$, $SiO_2$, $Na_2O$, S and $K_2O$ immediately adjacent to the mineralization; and removal of $SiO_2$ and $Na_2O$ and addition of $H_2O$, $CO_2$, and CaO in regions away from the mineralization. Trace elements that can be enriched adjacent to mineralization include gold, arsenic, antimony, barium, bismuth, lithium, mercury, rubidium, strontium, silver, sulphur, tungsten (Fyon and Crocket 1982; Kerrich 1983; Durocher 1983; Durocher and Hugon 1983; Robert and Brown 1984; Melling et al. 1986). The bulk composition of the rock, the fluid:rock ratio, and the ambient pressure and temperature conditions influence the type of alteration mineralogy (Colvine et al. 1988).

Generally, adjacent to mineralization in greenschist-facies tholeiite or komatiite (*see* Table 22.9), alteration is represented by ferroan carbonate and potassium-bearing mica (Fyon and Crocket 1982; Melling et al. 1986; Robert and Brown 1984; Piroshco and Hodgson 1988). Sericite is the mica in altered tholeiites and fuchsite occurs in komatiites. Biotite is more common in hydrothermally altered mafic and ultramafic rocks at upper greenschist facies (Andrews et al. 1986; Marmont 1986). Multiple, superimposed alteration assemblages can occur immediately adjacent to mineralized zones (Robert and Brown 1984; Piroshco and Hodgson 1988). Assemblages containing chlorite, actinolite, tremolite, talc and calcite are developed in altered tholeiites and komatiites farther away from the auriferous zone (Fyon and Crocket 1982; Melling et al. 1986; Piroshco and Hodgson 1988). Alteration associated with some of these deposits is superimposed upon regional greenschist assemblages (Robert and Brown 1984; Piroshco and Hodgson 1988).

Aluminous minerals (*see* Table 22.9) are spatially associated with hydrothermally altered mafic and felsic metavolcanic rocks in proximity to gold deposits in amphibolite-facies rocks (e.g., Madsen–Starratt–Olsen deposit and Campbell–A.W. White deposit, Red Lake camp—Durocher 1983, Andrews et al. 1986; Musselwhite prospect—Hall and Rigg 1986; Detour deposit—Marmont 1986; Dumagami deposit, Bousquet camp—Marquis et al. 1990a, 1990b; Hemlo deposit, Hemlo camp—Kuhns 1986, Muir et al. 1991). However, detailed observations in the Detour (Marmont 1986) and Dumagami deposits (Marquis et al. 1990a, 1990b), and the Campbell–A.W. White deposit (Andrews et al. 1986) indicate that alteration temporally related to gold, including chlorite, biotite and kaolinite, retrogrades the earlier aluminous alteration minerals. This indicates that the introduction of gold, and attendant hydrothermal activity, postdated peak metamorphism (Marmont 1986; Marquis et al. 1990a, 1990b).

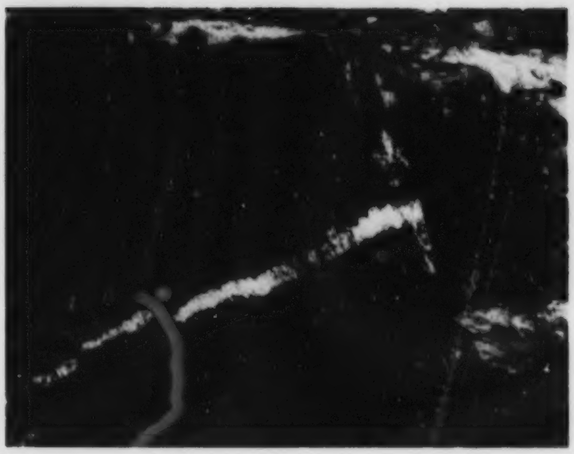

**Figure 22.16.** Quartz-carbonate-gold vein, Carshaw deposit (Gail Resources Inc.), Timmins. The vein cuts a banded, oxide-facies iron formation. A pyrite-gold replacement selvage is preferentially developed adjacent to the quartz vein in magnetite-rich mesobands.

Other general alteration types that are spatially and temporally associated with gold mineralization include silicification, sericitization, biotitization, sulphidation, feldspathization (both albite and microcline) (*see* Table 22.9). Iron-rich rocks, such as oxide-facies iron formation, are more susceptible to sulphidation, where oxide minerals (magnetite) are replaced by sulphide minerals (e.g., Figure 22.16, Carshaw–Malga deposit, Timmins–Porcupine camp—Fyon et al. 1983; Macleod–Cockshutt deposit, Geraldton camp—Matheson and Douglas 1948, Macdonald 1984; Musselwhite prospect—Hall and Rigg 1986; Central Patricia Mine, Pickle Lake camp—Ferguson 1966). Hydrothermal microcline occurs in deformed and altered felsic metavolcanic and metasedimentary rocks spatially associated with the Hemlo deposit (Kuhns 1986; Muir et al. 1991). Albite is also a common hydrothermal mineral found in auriferous quartz veins and immediately adjacent to mineralized zones, regardless of host rock composition (Hurst 1935; Keys 1940; Ebbutt 1948; Melling et al. 1986).

## STRUCTURAL CHARACTERISTICS OF MINERALIZED HIGH-STRAIN ZONES

Gold deposits occur along high-strain zones having specific strike orientations (*see* Table 22.9; Hodgson and Troop 1988): 1) northeast to east-northeast (e.g., Madsen–Starratt–Olsen deposit, Red Lake camp; Hollinger–McIntyre–Coniaurum deposit, Timmins–Porcupine camp; Holt–McDermott deposit, Harker–Holloway camp; Macassa deposit, Kirkland Lake camp; Kerr Addison–Chesterville deposit, Larder Lake camp; Pickle Crow–Central Patricia deposit, Pickle Lake camp); 2) east-southeast to southeast (e.g., Hemlo deposit, Hemlo camp; Campbell Red Lake-A.W. White deposit, Red Lake camp; Detour deposit, Detour camp). The thickness of these high-strain zones varies from centimetres to tens of kilometres, when considered on a stope to deposit scale (Robert 1990).

## Geometry of Deposits

Deposits are characteristically tabular, generally relatively narrow, and have much greater strike length and down-dip dimension, with respect to thickness. For example, the down-dip dimension to thickness aspect ratio of the Kerr Addison–Chesterville deposit, Larder Lake camp, is a minimum of approximately 16:1, and the down-dip dimension to strike length aspect ratio is about 1.5:1 (Smith et al. 1990). Similar ratios describe vein shoots in the Pickle Crow and Central Patricia deposits (Ferguson 1966).

## Strike of Mineralized High-Strain Zone, Deposit and Ore Shoots

Individual gold deposits and ore shoots within high-strain zones generally have strikes that are similar to those of the nearby subsidiary and regional structures (*see* Tables 22.8 and 22.9). However, in some deposits, differences exist between the strike of the host shear zone or local fold axial plane, and the strike of the deposit or ore shoots within the deposit (*see* Table 22.9). Strike differences are generally in the range of 10° to 30° (*see* Table 22.9). These slight differences between the strikes of the host high-strain zones and the deposits and ore shoots are consistent with the formation of vein and mineralization zones along subsidiary fractures within the high-strain zone, during the deformation event. Some of these fracture geometries are consistent with formation as Riedel shear fractures (Andrews et al. 1986).

## Plunge of Mineralized High-Strain Zone, Deposit and Ore Shoots

In most deposits, the plunge of the deposit or ore shoot is similar to that of stretch lineations in the host high-strain zone (*see* Tables 22.8 and 22.9). In the Geraldton camp, the plunge of the Hardrock–Macleod Cockshutt deposit is similar to that of regional asymmetric folds within the host regional-scale Barton Bay Deformation Zone. Plunges of discrete ore shoots within folded veins of the Pickle Crow–Central Patricia deposit are parallel to those of fold axes associated with the regional folds (Ferguson 1966). Deposits in the Shebandowan greenstone belt also plunge parallel to local stretch directions (Stott and Schnieders 1983). A deposit plunge may differ from the plunge of the contained ore shoots. For example, the Campbell–A.W. White deposit, Red Lake camp (Figure 22.17) plunges moderately to the southeast, whereas the discrete ore shoots within the deposit plunge moderately to the west and southwest (Rogers 1990; Figure 22.17). Similarly, the plunge of ore shoots in the Madsen and Starratt–Olsen mines is nearly orthogonal to that of the deposit (*see* Table 22.9; Horwood 1940).

The plane defined by the nearly orthogonal plunge directions of the deposits and ore shoots corresponds to the Y-X strain ellipse axis defined for the spatially associated high-strain zone (Andrews et al. 1986). One plunge corresponds to the stretch lineation (X-direction) and the other to the plane of maximum permeability (Y-direction) (H. Poulsen, Geological Survey of Canada, personal communication, 1991).

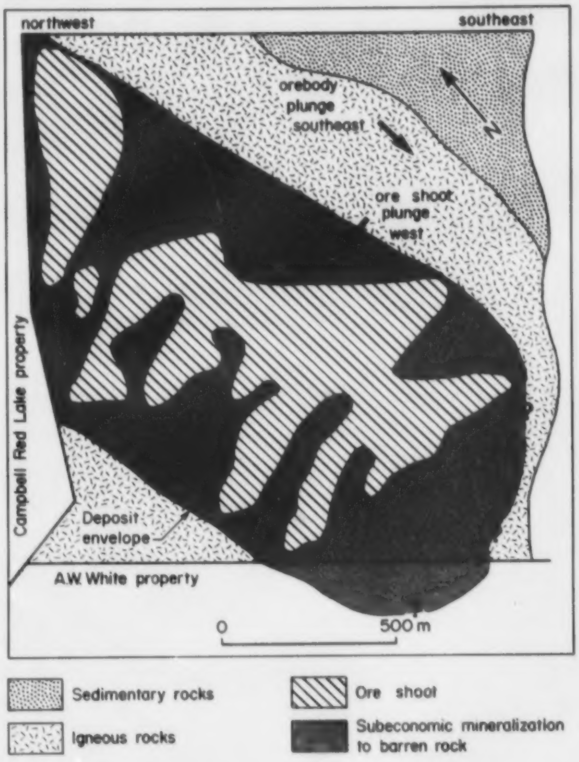

**Figure 22.17.** Schematic longitudinal projection of the East South "C" zone, in the A.W. White (Dickenson) deposit, Red Lake camp. Two contrasting plunges are observed. The deposit itself plunges moderately to the southeast. Within the deposit, ore shoots plunge moderately to the west and southwest (*modified after* Rogers 1990). The ore shoot plunge corresponds to the X- or stretching direction or the host high-strain zone. The deposit plunge corresponds to the Y-direction in the strain ellipse. The X-Y plane represents a zone of enhanced permeability.

## MINERALIZATION HABIT

Gold within mineralized zones occurs in several different settings, including quartz veins, sulphide-rich envelopes to quartz veins, and sulphide replacement zones which lack quartz veins or which are spatially removed from a quartz-vein filled fluid conduit. The development of a particular mineralization habit is partly dependent on the competency contrast and bulk composition of the host rocks. Competent rocks, such as basalt (Campbell Red Lake–A.W. White deposit, Red Lake camp—Andrews et al. 1986), granitoid or mafic dikes (e.g., albitized mafic dikes in the Kerr Addison-Chesterville deposit, Larder Lake camp—Smith et al. 1990), and iron formation units (e.g., Macleod–Cockshutt north deposit, Geraldton camp—Matheson and Douglas 1948, Macdonald 1984; Carshaw–Malga deposit, Timmins-Porcupine camp—Fyon et al. 1983; Musselwhite iron formation prospect—Hall and Rigg 1986; Pickle Crow–Central Patricia deposit, Pickle

Lake camp—Thomson 1939, Ferguson 1966), tend to fracture within high-strain zones and are usually the loci of quartz veins.

Bulk compositions which are rich in iron, such as oxide-facies iron formation or iron-rich tholeiite units, are also preferentially replaced by iron sulphide minerals, whereas iron-rich oxide or silicate minerals, such as magnetite or actinolite, are replaced by pyrite, pyrrhotite or arsenopyrite (e.g., Macleod-Cockshutt north deposit, Geraldton camp—Matheson and Douglas 1948, Macdonald 1984; Musselwhite iron formation prospect—Hall and Rigg 1986; Carshaw–Malga deposit—Fyon et al. 1983; Pickle Crow–Central Patricia deposit, Pickle Lake camp—Thomson 1939). More extensive replacement within iron formation units commonly results in the development of stratabound, auriferous, sulphide-rich zones which can extend well away from quartz veins (Figure 22.18).

## Quartz Vein Habit

Deposits comprising quartz veins are abundant in the Superior Province (*see* Table 22.9; Thomson 1939; Baker et al. 1957; Rogers 1982; Hodgson and MacGeehan 1982;

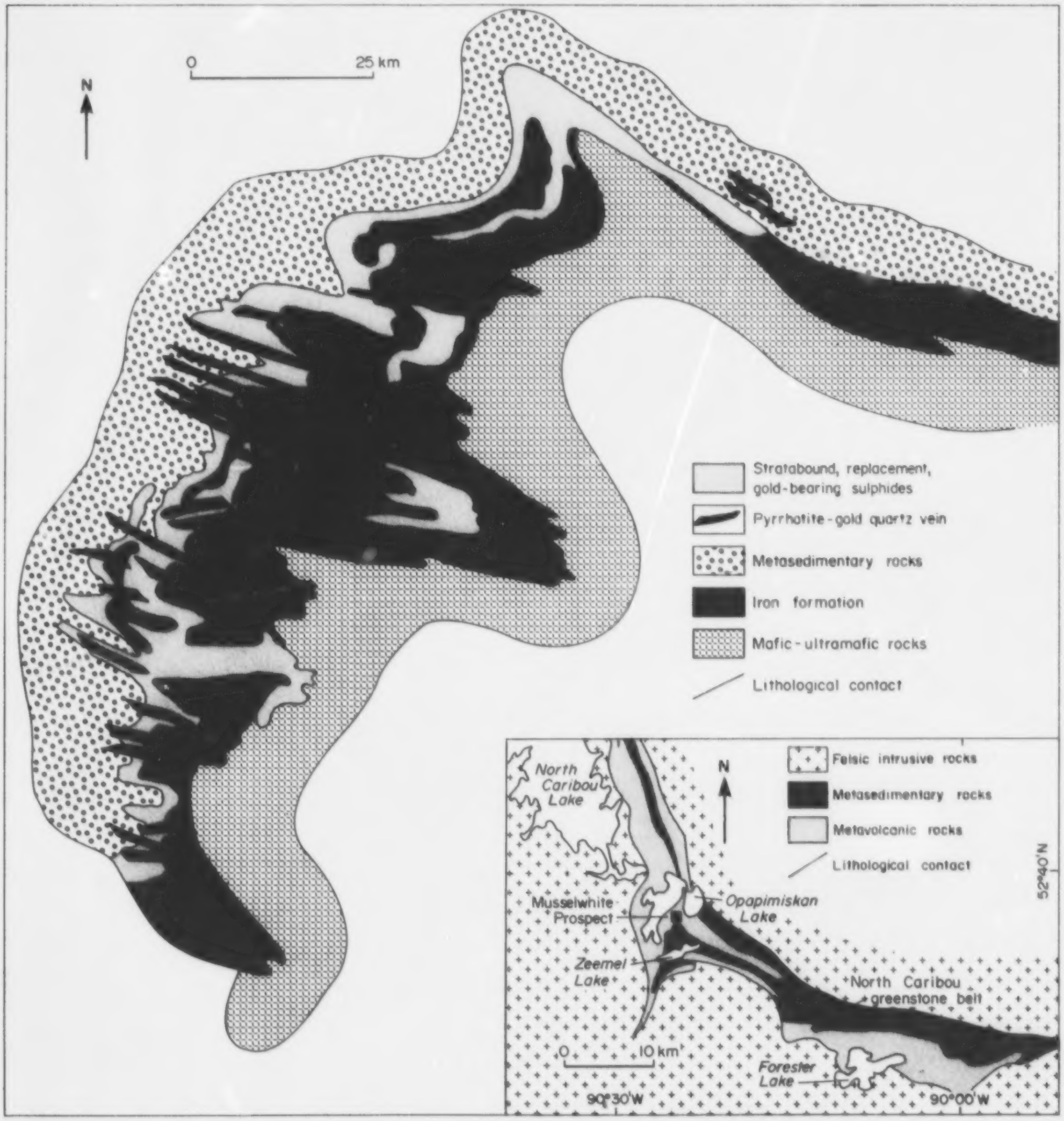
0
25 km
N
Stratabound, replacement, gold-bearing sulphides
Pyrrhotite-gold quartz vein
Metasedimentary rocks
Iron formation
Mafic-ultramafic rocks
Lithological contact
North Caribou Lake
N
Felsic intrusive rocks
Metasedimentary rocks
Metavolcanic rocks
Lithological contact
52°40'N
Musselwhite Prospect
Opapimiskan Lake
Zeemel Lake
North Caribou greenstone belt
0
10 km
Forester Lake
90°30'W
90°00'W

**Figure 22.18.** Localization of quartz veins along fracture cleavage, axial planar to northwest striking folds, in brittle, oxide-facies iron formation, Musselwhite prospect, North Caribou greenstone belt (*after* Hall and Riggs 1986). Vein-filled fractures did not propagate in to the adjacent metasedimentary and ultramafic rocks which were more ductile during folding.

Hodgson 1983; Wood, Burrows et al. 1986; Robert and Brown 1984, 1986; Kishida and Kerrich 1987). The auriferous veins generally consist of quartz and ferroan carbonate, accompanied by accessory albite, tourmaline, white mica and sulphide mineral veins (Ebbutt 1948). Vein-envelope alteration in mafic and ultramafic rock consists of sericite or fuchsite and ferroan carbonate (Bain 1933). Gold occurs as native grains and as inclusions in pyrite. Associated sulphide minerals can include pyrite with trace amounts of chalcopyrite, sphalerite, or galena (*see* Table 22.9; Keys 1940). In some deposits, the presence of galena, chalcopyrite or sphalerite signifies higher gold grade. Vein systems within, and spatially associated with, alkalic intrusions in the Kirkland Lake and Matachewan camps of the Abitibi greenstone belt contain notable amounts of telluride minerals (Burrows and Hopkins 1925; Charlewood 1964; McInnes et al. 1986).

Typically, the paragenetic sequence within the quartz veins is complex (Ebbutt 1948; Ferguson 1966; McInnes et al. 1986; Fyon 1986). Commonly, auriferous veins are characterized by multiple opening and filling events, and complex strain histories. The site of gold concentration in a single vein system can be quite variable, and can include, but may not be coincident with, specific alteration assemblages, sulphide concentrations, fractured or cataclastic vein quartz, specific rock types or specific structural features (*see* Table 22.10). In some areas, barren quartz veins do not display the same deformation history or deformation intensity as the auriferous arrays (Figure 22.19 ; Fyon and O'Donnell 1986; Heather 1989; Sage and Heather 1991). In other deposits, auriferous quartz veins retain delicate, primary depositional textures; some of these quartz vein arrays formed within, or along, ductile shear zones, by crack-seal mechanisms (Figure 22.20; Ramsay 1980) where increments of shear, resulting in dilation, were localized along vein margins (e.g., Sigma Mine—Robert and Brown 1986). In many deposits, some veins appear to have formed following most deformation. Thus, the controls on vein formation and gold precipitation are complex. Gold may not have coprecipitated with the vein quartz. Several generations of vein quartz may exist, only one of which coprecipitated with gold. In some deposits, some gold mineralization appears to have been introduced following the initial formation of the quartz vein, during a particular deformation event (e.g., Renabie deposit—Heather 1989).

A spectrum of rock types, including komatiite, tholeiite, wacke, oxide-facies iron formation, felsic dikes and small stocks are host to quartz veins (Hodgson and MacGeehan 1982). Individual veins can occur along shear zones or parallel to shear foliation (e.g., Sigma Mine—Robert and Brown 1986), parallel to fold axial surfaces (e.g., Musselwhite Prospect—Hall and Rigg 1986; *see* Figure 22.18), or parallel to shear fractures (e.g., Rouyn–Beauchastel area—Gauthier et al. 1990). In more competent rocks, veins can be oriented at high angles to rock layering (e.g., MacLeod-Cockshutt north deposit—Macdonald 1984).

## Disseminated Sulphide Habit

This habit of auriferous sulphide minerals (Figure 22.21) is prominent in the following amphibolite-grade deposits: Hemlo (Kuhns 1986; Muir et al. 1991); sulphidized, auriferous iron formation at the Musselwhite prospect (Hall and Rigg 1986); and the amphibolite-grade portion of the Campbell Red Lake–A.W. White deposit (e.g., arsenopyrite-quartz replacement zones and sulphide replacement bodies; Andrews et al. 1986; *see* Table 22.9); and in greenschist-facies rocks of the Central Patricia Mine (Thomson 1939). It is characterized by disseminated sulphide minerals, including pyrite, pyrrhotite and arsenopyrite (*see* Table 22.9). Although not necessarily prominent (e.g., Pickle Crow–Central Patricia deposit—Thomson 1939), quartz veins generally occur within the disseminated sulphide zones (Andrews et al. 1986). In the Musselwhite iron formation prospect, disseminated sulphide minerals extend laterally away from the quartz veins to define stratabound replacement zones within the iron formation (*see* Figure 22.18; Hall and Rigg 1986).

## Auriferous, Pyritic, Stratabound Deposits

Examples of this mineralization habit, also known as pyritic-type gold deposits (Robert 1990), include those in the Bousquet (e.g., Dumagami Mine—Marquis et al. 1990a, 1990b) and Casa Berardi–Joutel (e.g., Agnico-Eagle deposit) camps of Quebec. No economically viable Ontario examples are known; possible sites for this type of deposit are described later in this section. These deposits consist of stratabound, auriferous, massive pyrite lenses that occur within successions of deformed felsic metavolcanic rock. Quartz veins are present, but not abundant. The sulphide lenses, commonly base metal bearing, are interpreted to be syngenetic in origin and spatially associated with a premetamorphic aluminous alteration (Robert 1990; Marquis et al. 1990a, 1990b). In the Bousquet camp, gold-poor, pyritic massive sulphide deposits, having some characteristics of VMS deposits, can occur within or outside of the Dumagami structural zone, a subsidiary high-strain zone adjacent to the Larder-Cadillac deformation zone (Tourigny et al. 1988; Marquis et al. 1990a, 1990b). However, all auriferous massive sulphide lenses in the Bousquet camp occur within the Dumagami structural zone (Marquis et al. 1990b).

The origin of the deposits in the Bousquet camp has been the subject of much debate, and the complex tectonometamorphic history of the Bousquet district has led to 2 contrasting descriptive models: 1) syngenetic concentration of gold in exhalative sulphide lenses (Valliant et al. 1982; Stone 1990); and 2) gold superimposed upon syngenetic, but gold-poor, massive sulphide lenses during, or following, metamorphism (Tourigny et al. 1988; Marquis et al. 1990a, 1990b). Recent detailed structural analysis of the Bousquet area and the deposits indicates that (Marquis et al. 1990a, 1990b): 1) the auriferous lenses occur within the Dumagami structural zone, a high-strain zone subsidiary to the Larder–Cadillac deformation zone; 2) the auriferous lenses consist of gold-silver-copper-bearing sulphide minerals that were

**Table 22.10.** Structural features which have been attributed to the localization of, or which correspond to structural attitudes of, camps, gold deposits and ore shoots in parts of the Superior Province.

| Scale | Structural Characteristic | Example (Reference) |
|---|---|---|
| **Camp** | Intersection between northeast-striking segments of regional high-strain zones and northwest- to north-striking lineaments and faults. | Timmins, Ross, Kirkland Lake and Rouyn–Beauchastel camps, Abitibi greenstone belt (2,13). |
| | Northeast- to east-northeast-striking segments of regional deformation zones. | Timmins, Ross, Kirkland Lake and Larder Lake camps, Abitibi greenstone belt (13).<br>Madsen, Starratt–Olsen, Howey and Hasaga mines, Red Lake camp, Red Lake greenstone belt (9). |
| | Northwest- to west-northwest-striking segments of regional deformation zones. | A.W. White, Campbell Red Lake, H.G.Young, McMarmac, Cochenour–Willans, Gold Eagle and McKenzie mines, Red Lake camp, Red Lake greenstone belt (9).<br>Hemlo camp, Schreiber–Hemlo greenstone belt (6). |
| **Deposit** | High-strain zones created by, or reactivated during, the emplacement of plutons. | Deposits related to the Canoe Lake stock, Lake of the Woods area; Dona Lake deposit, related to the Ochig Lake pluton (14,15,17). |
| | Intersection between quartz vein systems and rock unit contacts. | Plunge of Pamour No. 1 and Hallnor deposits, Timmins–Porcupine camp, Abitibi greenstone belt (12). |
| | Intersection between steeply dipping high-strain zone and moderately dipping intrusions. | Plunge of Hollinger–McIntyre–Coniaurum deposit, Timmins–Porcupine camp, Abitibi greenstone belt (12). |
| | Intersection between late north- to northeast-striking diabase dikes and regional-scale ductile shear zones | Goudreau–Lochalsh and Missanabie–Renabie areas, eastern Michipicoten greenstone belt (5,7). |
| | Zones of cataclastic deformation superimposed upon ductile shear zones or quartz veins. | Dumagami Mine, Bousquet camp, Abitibi greenstone belt (3).<br>Pickle Crow Mine, Pickle Lake camp, Pickle Lake greenstone belt (16). |
| | Intersection between regional and subsidiary or sympathetic high-strain zones. | Cameron Lake deposit, Savant–Crow Lake greenstone belt (11). |
| | East-northeast-striking zones of z-shaped crenulation superimposed upon regional-scale shear foliation. | Rouyn–Beauchastel area, Abitibi greenstone belt (2).<br>Dumagami Mine, Bousquet camp, Abitibi greenstone belt (3). |
| | Hinges of late z- or s-shaped, asymmetrical folds of shear foliation, bedding, compositional layering or rock units. | Musselwhite prospect, North Caribou greenstone belt — folded oxide-facies iron formation (10).<br>Pickle Crow Mine, Pickle Lake camp, Pickle Lake greenstone belt (16,18).<br>MacLeod–Cockshutt mine, Geraldton camp, Beardmore–Geraldton greenstone belt — folded oxide-facies iron formation (4).<br>Renabie Mine area, eastern Michipicoten greenstone belt — folded high-strain zone (5,7).<br>Rouyn–Beauchastel area, Abitibi greenstone belt — folded rock units (2).<br>Camflo–Malartic Hygrade, Abitibi greenstone belt — folded rock units (8). |
| **Stope** | Fractures oriented at a low angle to local schistosity (possibly Riedel shear fractures and tension fractures within ductile shear zones). | Goudreau–Lochlash area, eastern Michipicoten greenstone belt (1).<br>Pickle Crow and Central Patricia mines, Pickle Lake camp, Pickle Lake greenstone belt (16,18).<br>Rouyn-Beauchastel area of Abitibi greenstone belt (2).<br>Foliation oblique ferroan carbonate veins in A.W. White Mine, Red Lake camp, Red Lake greenstone belt (9). |
| | Intersection between vein and host rock units, such as iron formation. | Pickle Crow Mine, Pickle Lake camp, Pickle Lake greenstone belt (18). |
| | Zones of east-southeast-striking crenulation superimposed upon east-striking shear foliation. | Rouyn–Beauchastel area, Abitibi greenstone belt (2). |

*References: 1– Heather and Arias 1987; 2– Gauthier et al. 1990; 3– Marquis et al. 1990a, 1990b; 4– Macdonald 1984; 5– Sage and Heather 1991; 6– Muir et al. 1991; 7– Heather 1989; 8– Sauve and Makila 1990; 9– Andrews et al. 1986; 10– Hall and Rigg 1986; 11– Melling et al. 1986; 12– Piroshco and Kettles 1987, 1988, 1991; 13– Hodgson and Troop 1988; 14– Smith and Thomas 1986; 15– Stott and Brown 1986; 16– Ferguson 1966; 17– Cohoon 1986; 18– Thomson 1939.*

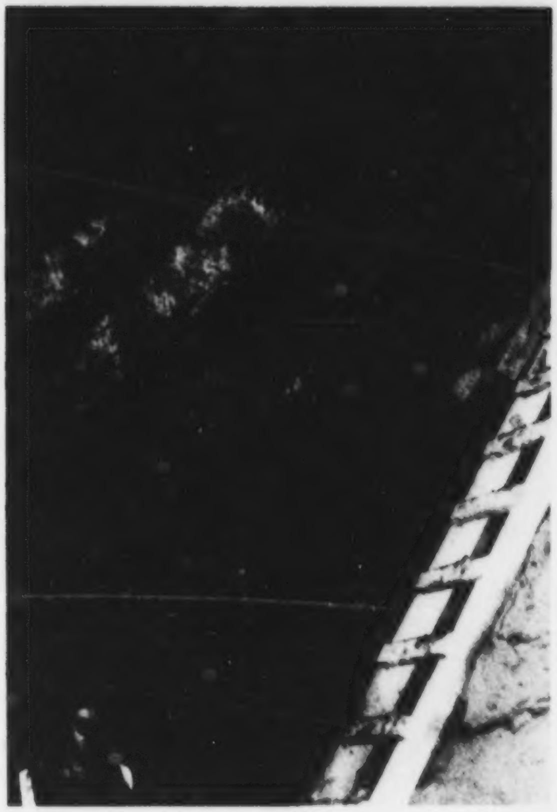

**Figure 22.19.** Banded, crack-seal gold-quartz-carbonate vein (No. 51 vein), 1400 level, Pamour No. 1 Mine (Royal Oak Mines Inc.), Timmins. This habit of gold-bearing quartz vein is typical of several deposits in the Superior Province.

superimposed upon deformed massive sulphide lenses following peak metamorphism and ductile deformation; and 3) the auriferous zones within the pyritic lenses are coincident with zones of late cataclastic deformation and attendant retrograde alteration. These are interpreted to be syngenetic VMS deposits that were deformed in a gold-associated subsidiary high-strain zone, and contain a superimposed component of gold mineralization (Marquis et al. 1990a, 1990b).

Other gold-rich VMS deposits that may have experienced a similar history and contain a superimposed component of gold mineralization, include the Horne, Quemont, and Deldona VMS deposits in the Noranda camp (K.B. Heather, Ontario Geological Survey, personal communication, 1991) and the Joutel deposit in the Casa Berardi camp (Simard and Genest 1990). The auriferous VMS deposits in the Noranda area occur in proximity to the east-northeast-striking Horne Fault, interpreted to represent the structural margin of the Noranda Cauldron (Gibson and Watkinson 1990). The structural history of the Horne Fault is not known in detail, but late increments of motion cannot be ruled out (K.B. Heather, Ontario Geological Survey, personal communication, 1991). The Joutel deposit likewise occurs in proximity to VMS deposits (Joutel camp), along a southeast-striking deformation zone called the "Harricana fault" (Simard and Genest 1990). This style of mineralization has not been widely recognized in Ontario, although the Hemlo deposit has been described as this type (Valliant and Bradbrook 1986). Rock successions in Ontario that have the potential to host this type of deposit include assemblages which are known to contain syngenetic pyritic lenses (VMS or iron formation units), that are located within, or in proximity to, regional or subsidiary deformation zones known to be associated with gold mineralization. In the Abitibi greenstone belt, prospective areas include the southwestern part of the Blake River assemblage (Jackson and Fyon, this volume), north of the Kirkland Lake–Virginia Town area, along the Larder–Cadillac "break", and assemblages in the Burntbush area (Jackson and Fyon, this volume) along the Casa Berardi "break" (Pattison et al. 1986), Abitibi greenstone belt.

### Mineralization Habit and Metamorphic Grade

Suggestions have been made that the quartz vein habit is not characteristic of deposits that formed under amphibolite-grade metamorphic conditions (Colvine et al. 1988). Such an inference was based partly on: 1) the observation that, world-wide, most quartz vein deposits occur in rocks of greenschist metamorphic grade (Woodall 1979); and 2) the change from predominantly a vein habit to a disseminated habit across the greenschist-amphibolite metamorphic isograd in the Campbell Red Lake–A.W. White deposit, Red Lake camp (Andrews et al. 1986).

Most, if not all, of the gold deposits that occur in rocks of amphibolite metamorphic grade in the Superior Province contain a quartz vein component (Andrews et al. 1986; Marmont 1986; Hall and Rigg 1986; Marquis et al. 1990a, 1990b; Muir et al. 1991). Gold was introduced following peak metamorphism, and is associated with a "retrograde" alteration assemblage in some of these deposits (e.g., Dumagami—Marquis et al. 1990a, 1990b; Detour—Marmont 1986). Also, in most deposits, regardless of ambient metamorphic grade, a variety of mineralization habits exist, including quartz vein and disseminated sulphide (e.g., A.W. White quartz vein, replacement ore and mineralized sulphide ore—Rogers 1990; Kerr Addison pyritic flow ore, green carbonate quartz vein ore, and albitite quartz vein ore—Smith et al. 1990).

These observations indicate that a correspondence may exist between mineralization habit and ambient metamorphic grade of host rocks, but the constraints on mineralization habit are complex and are not simply the result of ambient temperature and pressure conditions.

## Descriptive Classification of Superior Province "Gold-Only" Deposits

Gold deposits in the southeastern Abitibi Subprovince in Quebec have been subdivided into 2 deposit classes (Robert 1990): 1) pyritic, and 2) quartz vein. The presence of attributes of one class within deposits nominally assigned to the other class causes us to shy away from use of this scheme.

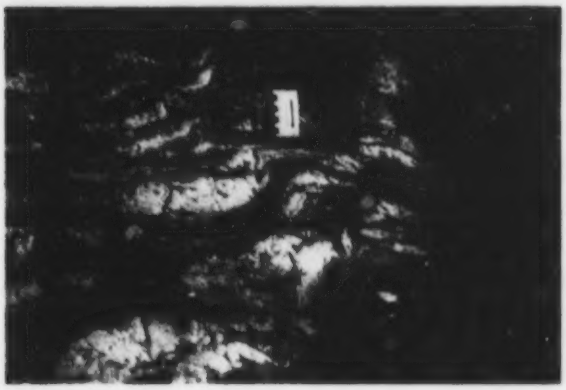

**Figure 22.20.** S-folded quartz veins in a highly strained tonalite, Renabie Mine area. In this area, those quartz veins that contain higher grades and larger masses of gold are more highly strained. However, the style and relative age of deformation features also influence gold tenor. (Photo *courtesy of* K.B. Heather, Ontario Geological Survey).

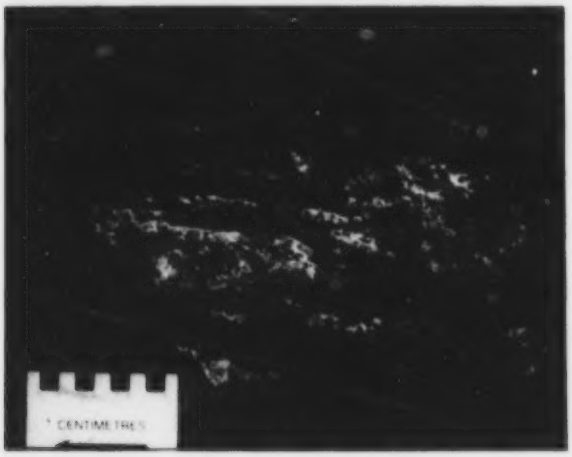

**Figure 22.21.** Gold mineralization from the Hemlo ore deposit. The mineralization consists of disseminated pyrite and molybdenite in a feldspathized host rock. Note the folded and dismembered quartz veins, which generally constitute a minor part of the mineralized rock. (Photo *courtesy of* T. Muir, Ontario Geological Survey).

For example, quartz veins occur in fractured competent rocks in most deposits. Hence, the presence or absence of quartz veins is not a reliable discrimination criterion. Similarly, disseminated, replacement sulphide minerals are common in sulphidized oxide-facies iron formations (Thomson 1939; Ferguson 1966; Macdonald 1984; Fyon et al. 1983; Hall and Rigg 1986) as envelope-alteration zones adjacent to quartz veins, as foliation-parallel zones within ductile shear zones, associated with some deformed quartz veins (Andrews et al. 1986) and superimposed upon syngenetic pyritic bodies (e.g., Bousquet district—Marquis et al. 1990a, 1990b). Hence, this habit may not provide a reliable discrimination. Indeed, the presence of more than one habit of mineralization is commonly observed within a single deposit or stope (Baker et al. 1957; Thomson 1939; Kishida and Kerrich 1987; Smith et al. 1990; Rogers 1990).

## Localization of Deposits Within High-Strain Zones

The structural history of a high-strain zone determines the existence and geometry of permeable zones required to channel hydrothermal fluids. It is not surprising, therefore, that the geometry of the deposit and ore shoots are similar to the geometry of structural elements within host high-strain zones (*see* Strike of Mineralized High-Strain Zone, Deposit and Ore Shoots and Plunge of Mineralized High-Strain Zone, Deposit and Ore Shoots, above; *see* Table 22.10). The position of some deposits along, and within, the host high-strain zone are also partly determined by the intersection of structural elements.

### LOCATION OF CAMPS AND DEPOSITS ALONG HIGH-STRAIN ZONES

The Timmins–Porcupine, Ross and Kirkland Lake camps, in the Ontario portion of the southwestern Abitibi greenstone belt, appear to be localized where northwest-trending lineaments intersect the Porcupine–Destor and Larder–Cadillac deformation zones (Hodgson 1986). Several camps are localized along east-northeast-striking segments of regional deformation zones (*see* Table 22.8). Some deposits (e.g., Renabie—Heather 1989, Sage and Heather 1991; Campbell Red Lake–A.W. White deposit—Rogers 1990) are localized along a portion of a high-strain zone that was transposed into an east-southeast orientation. This commonly occurring orientation is also observed in the Sachigo Subprovince, where mineralization at the Musselwhite prospect is localized along, or parallel to, an east-southeast-striking cleavage, which is axial planar to folds within iron formation (Hall and Rigg 1986).

Other deposits, such as the Macleod–Cockshutt (Matheson and Douglas 1948; Macdonald 1984), and deposits in the Rouyn–Beauchastel area (Gauthier et al. 1990), are localized in segments of a high-strain zone where compositional layering is asymmetrically folded. At the Macleod–Cockshutt deposit, mineralization decreases in grade and intensity down-plunge to the west, where the folded iron formation becomes more open (Macdonald 1984). The Dona Lake deposit, Pickle Lake greenstone belt (Stott and Brown 1986; Cohoon 1986) and those spatially associated with the Canoe Lake stock in the western Wabigoon greenstone belt (Smith and Thomas 1986), occur in high-strain zones that were created or reactivated by the emplacement of plutons (Stott and Smith 1988).

### CORRESPONDENCE BETWEEN GEOMETRIES OF DEPOSITS AND STRUCTURAL FEATURES

The plunge of some deposits and their ore shoots can correspond to several structural elements, including intersection lineations between structural elements, or stretch lineations within a high-strain zone. The location and plunge of gold-bearing shoots in the Cameron Lake deposit, Savant Lake–Crow Lake greenstone belt, corresponds, in part, to the interaction between the Cameron Lake deformation zone and a set of sympathetic high-strain zones (Melling et al. 1986). In the Timmins–Porcupine camp, the east-northeast plunge of the Hollinger–McIntyre deposit corresponds to the intersection between high-strain zones and porphyry bodies (Piroshco and Kettles 1991). Also, in the Timmins–Porcupine camp, the trend and plunge of deposits along, and adjacent to, the unconformity between metasedimentary and metavolcanic rocks in the northeastern part of the camp, corresponds to the intersection between moderately dipping quartz arrays and steeply dipping rock contacts (Piroshco and Kettles 1991). A similar control is illustrated in the Pickle Crow Mine, Pickle Lake camp, where the trend and plunge of the intersection lineation between iron formation and east northeast-striking quartz veins correspond to the trend and plunge of ore shoots (*see* Table 22.10; Thomson 1939).

Deposits and ore shoots in the Bousquet camp, Quebec, occur within syngenetic pyritic sulphide bodies that are cut by superimposed zones of cataclastic deformation (Tourigny et al. 1988; Marquis et al. 1990a, 1990b). A spatial correspondence also occurs between folded and cataclastized parts of quartz veins and ore shoots in the Pickle Crow Mine, Pickle Lake camp (*see* Table 22.10; Thomson 1939; Ferguson 1966). In the Rouyn–Beauchastel area, Quebec, auriferous zones occur where zones of east-southeast-striking crenulation are superimposed upon the Larder–Cadillac deformation zone (Gauthier et al. 1990).

Some ore shoots in ferroan carbonate veins in the Campbell Red Lake Mine, Red Lake camp, occur at shallow angles to the dominant shear foliation. The attitudes of these low-angle veins are consistent with their formation as Riedel shear fractures within high-strain zones (Andrews et al. 1986). A Riedel shear explanation has also been proposed to account for the variation in high-strain zone orientation in the northeastern part of the Michipicoten greenstone belt (Heather and Arias 1987); however, more recent interpretations of high-strain zone orientations in this part of the Michipicoten greenstone belt suggest a more complicated chronology than can be accounted for by Riedel shear fractures (Heather 1989; Sage and Heather 1991). Quartz veins in the Pickle Crow–Central Patricia deposit, Pickle Lake camp are also oriented at a shallow angle to regional foliation and may indicate the formation of veins by filling of brittle fractures (Riedel?) developed during folding (Thomson 1939).

## Age of Gold Deposits

The relative age relationships of gold introduction are constrained by: 1) cross-cutting relationships between gold-bearing structures or mineralization, and dated unmineralized intrusions; 2) deflection of gold-bearing structures by dated unmineralized intrusions; 3) the ages of hydrothermal zircons, titanites, rutile and micas collected from altered wall rock and auriferous veins. Most, but not all, structures which are host to gold generally developed, or were reactivated, during regional-scale deformation events. Formation of many gold deposits in the Superior Province of Ontario can be attributed to structural and hydrothermal events related to the Kenoran Orogeny. The hydrothermal activity spatially and genetically associated with the Kenoran gold deposits was protracted, and could have continued for at least 100 million years following the youngest Archean magmatism. This hydrothermal activity recorded in, and adjacent to, many Kenoran gold deposits was coeval with, and may have been related to, the thermal activity recorded in the middle crust (Card et al. 1989; Colvine 1989; Fyon et al. 1989). However, relationships in the Pickle Lake camp, Uchi Subprovince, and North Caribou greenstone belt, Sachigo Subprovince, are consistent with the existence of pre–Kenoran gold deposits as well. Thus, some evidence exists in support of multiple periods of gold deposit formation.

### GOLD RELATED TO OROGENY AT 2.86 Ma, SACHIGO SUBPROVINCE

Gold at the Musselwhite prospect, North Caribou greenstone belt, occurs in folded, sulphidized iron formation (Breaks, Bartlett et al. 1985; Hall and Rigg 1986). Gold occurs in quartz veins that occupy the fold axial traces and axial planar cleavage related to west-northwest-striking folds (*see* Figure 22.18). These folds formed in response to north-northeast-directed shortening, possibly related to the emplacement of the "late-tectonic", crescentic, 2.86 billion-year-old (D. Davis, Jack Satterly Laboratory, Royal Ontario Museum, unpublished data) North Caribou batholith, which was emplaced late in the tectonic history of the area (Stott et al. 1989).

If the introduction of gold into quartz veins and sulphide replacement zones along fold hinges was synchronous with folding of the iron formation (Hall and Rigg 1986), then gold introduction was approximately synchronous with the emplacement of the North Caribou batholith (2.86 Ga) during an orogeny which predated the Kenoran Orogeny by about 180 million years. However, the southeast-striking North Caribou Lake–Totogan Lake shear zone (Osmani and Stott 1988) passes within 2 km of the Musselwhite prospect. The age of this structure is presently not known, but it may have developed, or been reactivated, during the Kenoran Orogeny. Hence, gold at the Musselwhite prospect may alternatively have been introduced during the Kenoran Orogeny, approximately 300 million years after the emplacement of the 2.86 billion-year-old, posttectonic North Caribou batholith.

### GOLD DEPOSITS EMPLACED BEFORE 2741 Ma

In the Pickle Lake camp, gold deposits may have formed synchronously with (Dona Lake deposit), and prior to (Pickle Crow–Central Patricia deposit), the emplacement of 2741 million-year-old plutons (Stott, this volume). The Pickle Crow–Central Patricia deposit occurs within a northeast-trending structural domain characterized by northeast- to east-northeast-striking fold axes and foliation and north-northeast-plunging stretch lineations (Domain 1, Stott, this volume). The zone of mineralization extends into the adjacent Domain 3 (Stott, this volume). The Pickle Crow-Central Patricia gold deposit and a northeast-striking regional syncline are interpreted to have formed during a northwest-directed shortening event (Thomson 1939; Ferguson 1966; Stott, this volume). Stretch lineations, fold axes and axial planar foliation within the gold-bearing southwestern part of Domain 1 are deflected by the "late kinematic" Hooker–Burkoski stock, inferred to be approximately 2741 million years old (Stott, this volume). Thus, the deflected fabrics in Domain 1 formed prior to the emplacement of the Hooker–Burkoski stock.

If the Pickle Crow–Central Patricia gold deposit formed during the deformation event that produced the regional, northeast-striking folds (Thomson 1939; Ferguson 1966), then the gold deposit may also have formed prior to the emplacement of the 2741 million-year-old intrusion (Stott, this volume). However, the presence of the 2693 million-year-old, crescentic Second Loon pluton that was emplaced to the southeast of the Pickle Crow–Central Patricia gold deposit, is a complicating factor in the interpretation of the deposit age. Related to the Second Loon pluton is a contact strain aureole (Domain 3; Stott, this volume) that affects the strikes of fold axes and foliations, the trends and plunges of stretch lineations in Domain 1, and the gold mineralization at the Pickle Crow–Central Patricia deposit. The presence of a contact strain aureole related to a Kenoran intrusion opens the possibility that quartz veins in the Pickle Crow–Central Patricia gold deposit experienced a second deformation during the Kenoran Orogeny, at which time gold could have been introduced. Such a deformation may be manifested as Z-shaped, folded and cataclastized auriferous quartz veins in the Pickle Crow–Central Patricia deposit. These deformed zones within the quartz veins are the loci of gold mineralization (*see* Table 22.9; Thomson 1939; Ferguson 1966). Although these Z-shaped minor folds in the quartz veins have trends and plunges similar to those developed along the regional syncline (trend 020°; plunge 70°), the possibility still exists that all or some of the gold in this part of the Pickle Lake camp could have been superimposed upon a pre-existing quartz vein system during the Kenoran Orogeny.

## GOLD DEPOSITS EMPLACED AT APPROXIMATELY 2741 Ma

The Dona Lake gold deposit in the southern part of the Pickle Lake camp occurs within sulphidized and folded oxide-facies iron formation (Cohoon 1986). This iron formation occurs within a contact strain and metamorphic aureole developed adjacent to the Ochig Lake pluton (2741 Ma; Corfu and Stott 1989). The orientation of shear foliation, fold axes and stretch lineations within the gold deposit are similar to those of fold axes, stretch lineations and foliation that define the contact strain aureole adjacent to Ochig Lake pluton (Stott, this volume). No other fabrics occur in this area that can be attributed to a tectonic event younger than the emplacement of the Ochig Lake pluton (Stott, this volume). Hence, deformation of the mineralized iron formation took place in response to the emplacement of this pluton. If sulphidation of the iron formation and gold introduction accompanied this deformation event, then the data are consistent with the formation of the Dona Lake gold deposit prior to the Kenoran Orogeny, synchronously with the emplacement of the Ochig Lake pluton (2741 Ma; Stott 1991, this volume).

## KENORAN (2700 TO 2660 Ma) OR YOUNGER DEPOSITS

In the Uchi, Wabigoon, Wawa and Abitibi subprovinces, shear foliation, faults and fold axes within some regional deformation zones developed during Superior Province-scale, north-directed compression and dextral transpression (Stott 1986; Robert 1989; Hodgson et al. 1990; Gauthier et al. 1990; Marquis et al. 1990a, 1990b; Williams, Stott, Thurston et al., this volume). Most gold deposits in the Superior Province of Ontario are spatially associated with these regional deformation zones and appear to have formed during, or following, the Kenoran Orogeny.

In the Red Lake greenstone belt, Uchi Subprovince, calc-alkalic volcanism and contemporaneous plutonism had terminated by 2730 Ma (Corfu and Andrews 1987). Late, postvolcanic plutonism continued until *circa* 2700 Ma (Corfu and Andrews 1987; Nunes and Thurston 1980). Deformation, alteration and attendant gold mineralization in the Red Lake greenstone belt took place between approximately 2720 Ma, the age of the gold-mineralized McKenzie granodiorite and *circa* 2700 Ma, following regional greenschist-facies and contact amphibolite-facies metamorphism (Corfu and Andrews 1987). The presence of an altered 2714 million-year-old (Corfu and Andrews 1987) quartz feldspar porphyry dike, cross-cutting gold mineralization, may constrain the introduction of gold to between 2720 and 2714 Ma in the Campbell Red Lake–A.W. White deposit (Corfu and Andrews 1987).

In the Lake of the Woods area, Wabigoon Subprovince, alteration spatially related to gold mineralization is superimposed upon amphibolite-facies contact metamorphism related to a stock dated at 2708 ± 7 Ma (Davis and Smith 1991). Gold is interpreted to have been introduced into structures that developed during plutonism between 2716 and 2708 Ma (Davis and Smith 1991).

In the Shebandowan greenstone belt, western Wawa Subprovince, gold occurs in structures that had been activated by 2690 Ma (Corfu and Stott 1986; Stott et al. 1987). In the Hemlo camp, hydrothermal activity, possibly related to gold introduction, was most intense between 2690 and 2670 Ma (Corfu and Muir 1989). However, 2671 million-year-old $^{40}Ar/^{39}Ar$ ages for hydrothermal white micas (Masliwec et al. 1986), and a range of ages from 2650 to 2610 Ma for hydrothermal monazite and rutile (Corfu and Muir 1988) from the auriferous zones, indicate that the Hemlo deposit area experienced a protracted thermal and hydrothermal history.

In the southern Abitibi greenstone belt and in the northeastern Michipicoten greenstone belt, structures which localize gold deposits are superimposed upon, and therefore younger than, shear foliation and folds related to the regional-scale deformation zones that developed during the accretion of the Superior Province (Robert 1989; Hodgson et al. 1990; Gauthier et al. 1990; Marquis et al. 1990a, 1990b; Heather 1989; Sage and Heather 1991). Therefore, these gold-bearing structures, and by implication the contained gold deposits, formed late, during, or following, the Kenoran Orogeny. Geochronological data are also consistent with this empirical observation. In the Timmins–Porcupine camp, gold-bearing quartz veins cut 2691 to 2688 million-year-old (Corfu et al. 1989; Marmont and Corfu 1989) granodioritic porphyry intrusions and an albitized, 2673 million-year-old (Marmont and Corfu 1988) lamprophyre dike (Wood, Burrows et al. 1986; Marmont and Corfu 1988). The presence of 2633 to 2615 million-year-old hydrothermal sericite ($^{40}Ar/^{39}Ar$) within, and immediately adjacent to, auriferous veins in the Dome, Hollinger and Davidson–Tisdale mines (Masliwec et al. 1986; Hanes et al. 1987), suggests that the auriferous hydrothermal systems also experienced a protracted thermal history that extended for at least 60 million years following the last period of Archean magmatism. In the Val d'Or camp, Quebec, ages of hydrothermal rutile from the Sigma Mine (Wong et al. 1989; Jemielita et al. 1990), hydrothermal micas from the Sigma Mine (Wong et al. 1989), and scheelite (Bell et al. 1989) from alteration assemblages and from auriferous quartz veins, indicate that hydrothermal activity and attendant auriferous quartz veins may be as young as 2592 to 2579 million years old, having formed approximately 90 million years after the last upper crustal Archean magmatism.

## EPISODIC VERSUS DIACHRONOUS FORMATION OF GOLD DEPOSITS

The formation of gold deposits in the Superior Province was both episodic and diachronous. Gold deposits appear to have formed during 2 types of orogenic events: 1) construction or assembly of the granite-greenstone belts; and 2) large-scale accretion of superterranes to form the Superior Province. Evidence in support of gold deposit formation during the construction or assembly of granite-greenstone belts is still limited, but is preserved in windows within parts of the Uchi Subprovince that were not subjected to Kenoran orogenic events.

### Deposits Formed During Volcano-Plutonic Arc Construction or Assembly of Granite-Greenstone Belts

The Pickle Crow–Central Patricia gold deposit in the Pickle Lake camp may have formed during an orogenic event that took place before 2741 Ma. Possible orogenic events include the construction or juxtaposition of the North Caribou terrane and the Woman assemblage (before 2.77 Ga), or events related to early (*circa* 2740 Ma) construction of the Confederation and St. Joseph assemblages, to the south of the Pickle Lake camp (Stott, this volume). The Dona Lake gold deposit appears to have formed at approximately 2740 Ma, contemporaneously with the Ochig Lake pluton. This pluton was coeval with subduction-related volcano-plutonic arc construction taking place to the south in the Confederation and St Joseph assemblages (2740 to 2710 Ma; Stott, this volume).

### Deposits Formed During Province-Scale Accretion

Gold deposits formed during the waning periods of at least 2 regional-scale accretionary orogenies: 1) an orogeny which affected the northern Sachigo Subprovince *circa* 2.87 to 2.85 Ga; and 2) during or following the Kenoran Orogeny (2.72 to 2.60 Ga) in the Uchi, Wabigoon, Wawa and Abitibi subprovinces. Deposits that formed during, or following, the Kenoran Orogeny occur in the Uchi (Red Lake camp), Wabigoon (Lake of the Woods area), Wawa (Hemlo camp) and Abitibi (Detour, Timmins–Porcupine, Kirkland Lake, Val d'Or, Malartic and Bousquet camps) subprovinces and have been addressed previously (*see* section entitled Kenoran (2700 to 2660) or Younger Deposits).

The termination of volcanism and the onset of folding and thrusting related to the Kenoran Orogeny display a progressive southeastward younging from the Uchi Subprovince to the Abitibi Subprovince (Stott et al. 1987; Williams, Stott, Thurston et al., this volume). The approximate age of gold deposit formation in the Uchi, Wabigoon, Wawa and Abitibi subprovinces similarly decreases from *circa* 2720 to 2715 Ma in the western Uchi and Wabigoon subprovinces (Corfu and Andrews 1987; Davis and Smith 1991), to *circa* 2680 to 2670 Ma in the Hemlo deposit, central Wawa Subprovince (Corfu and Muir 1989), to *circa* 2673 to 2580 Ma in the southern Abitibi Subprovince (Masliwec et al. 1986; Hanes et al. 1987; Stott et al. 1987; Marmont and Corfu 1988; Colvine et al. 1988; Colvine 1989; Card et al. 1989; Wong et al. 1989; Fyon et al. 1989; Card 1990; Jemielita et al. 1990). Within the Abitibi greenstone belt, a similar north-to-south younging of gold deposit formation may also exist, but requires further documentation. In the northern part of the Abitibi subprovince, the maximum age of gold mineralization in the Detour Mine is $2722^{+3}_{-2}$ Ma (Marmont and Corfu 1989). However, the minimum age is not constrained, in contrast to 2673 to 2580 Ma for deposits along the southern part of the belt. These age relationships indicate that gold deposit formation in the Uchi, Wabigoon, Wawa and Abitibi subprovinces may have been diachronous during, or following, the Kenoran Orogeny.

## COMPARISON WITH UPPER AND MIDDLE CRUSTAL THERMAL RECORDS

The period of protracted hydrothermal activity associated with gold deposits ( from earlier than 2673 Ma to 2580 Ma) in upper crustal deformation zones continued for at least 70 million years after the youngest, upper-crustal plutonism and coeval contact metamorphism. This period of protracted, upper-crustal hydrothermal activity was coeval with the *circa* 2700 to 2630 million-year-old peak metamorphic and post peak-metamorphic thermal histories recorded in the middle crust exposed in the Kapuskasing Structural Zone and Winnipeg River subprovince ( Krogh et al. 1988; Corfu 1988). The apparent coeval hydrothermal age of both the upper and mid- to lower-crustal hydrothermal and metamorphic systems infers that the systems may have been coupled (Fyon et al. 1984; Card et al. 1989; Colvine 1989; Fyon et al. 1989), consistent with the crustal-scale processes that characterized the Kenoran Orogeny.

### Tectonic Setting

The Pickle Crow–Central Patricia and Dona Lake gold deposits may represent a preserved Archean analogue to Miocene to Quaternary gold deposits of the western Pacific island arcs. Gold metallogeny in the western Pacific was critically dependent on subduction of oceanic crust and the consequent construction of volcano-plutonic arcs (Sillitoe 1989). The formation of the Pickle Crow–Central Patricia and Dona Lake gold deposits appears to have been coeval with, and constrained by, magmatic and structural events related to volcano-plutonic arc construction or collision of terranes along the southern edge of the North Caribou and Pickle terranes to form the Uchi Subprovince. Terrane accretion in the northern part of the Uchi Subprovince took place around 2740 Ma, at least 20 million years prior to the onset of the Kenoran Orogeny. In this regard, the Pickle Crow–Central Patricia gold deposit possesses an alteration assemblage (dominantly chlorite in basalt) that is atypical of most Archean gold deposits formed in comparable rocks at comparable temperature and pressure conditions (*see* Table 22.9), perhaps indicative of different genetic processes.

Most of the Archean gold deposits in the Superior Province of Ontario appear to have formed during the late stages of the Kenoran Orogeny. The apparent north-to-south decrease in age of Kenoran gold deposits in the southern part of the Superior Province of Ontario requires further rigorous documentation and testing, but this trend is consistent with the documented north-to-south younging of volcanic, magmatic and structural events across the Uchi, English River, Wabigoon, Quetico, Wawa and Abitibi subprovinces (Williams, Stott, Thurston et al., this volume). This north-to-south younging is attributed to the final subduction-related accretion of superterranes that led to the construction of the Superior Province during the Kenoran Orogeny (Williams, Stott, Thurston et al., this volume). This geological record is similar to that of modern arc-arc accretion phenomena (Sillitoe 1989; Williams, Stott, Thurston et al., this volume). In southeastern Alaska, asymmetric, westward migrating fronts of volcanic, sedimentary, magmatic, structural and mineralization-forming events are observed across a series of tectono-stratigraphic terranes that were accreted to the west coast of North America during the Cretaceous to Eocene (Engebretson et al. 1985; Goldfarb et al. 1988). In the Juneau gold camp, the products of hydrothermal alteration and gold mineralization were being superimposed upon tectonized and metamorphosed sedimentary and igneous rocks during a period of regional uplift, simultaneously with the subduction and accretion of oceanic crust and related sedimentary rocks, 100 km to the west (Goldfarb et al. 1988). The development of mesothermal gold deposits within modern accretionary margins, in response to the subduction of hydrous oceanic crust, may be an analogue for Archean gold deposits that formed during, or following, the Kenoran Orogeny in the Superior Province of Ontario (Hodgson and Hamilton 1988).

## TECTONIC SYNTHESIS

A tectonic synthesis of the Superior Province within Ontario is summarized by Williams et al. (this volume). This tectonic history can be subdivided into 6 evolutionary stages (Figure 22.22): 1) formation of greenstone assemblages in disparate tectonic settings; 2) aggregation of greenstone assemblages into linear belts, on which are superimposed calc-alkalic volcanic centres and batholith complexes; 3) erosion of emergent volcanic material to fill prismatic sedimentary basins; 4) progressive crustal-scale shortening, attendant underthrusting of greenstone subprovinces beneath sediment prisms, and related tectonic thickening; 5) localized melting of thickened crust; and 6) continued shortening by a combination of underthrusting and lateral, subprovince-parallel slip. Some of these tectonic stages are associated with the formation of metallic mineral deposits.

In this section, we speculate on the relationships between the formation of specific metallic mineral deposits, Their time and space relationships to specific tectonic elements, including assemblage types, tectonic settings and structures (*see* Figure 22.22). Modern tectonic analogues are identified where this can be done with confidence.

### Deposits Formed During the Construction of Greenstone Belts

Volcanic assemblages which constitute greenstone belts in the Superior Province formed during at least 2 periods. Between 3.03 and 2.80 Ga, platform (Wood, Thurston et al. 1986; deKemp 1987; Thurston et al. 1987; Thurston and Chivers 1990) and mafic plain komatiitic assemblages, and spatially associated assemblages that contain minor felsic metavolcanic rocks and calc-alkalic plutons, developed and are preserved in parts of the Sachigo, Uchi, Wabigoon and Wawa subprovinces (Wilks and Nisbet 1985; Wood, Thurston et al. 1986; Thurston et al. 1987). The second period of greenstone belt construction took place between approximately 2.75 and 2.70 Ga in the Uchi, Wabigoon, Wawa and Abitibi subprovinces, during which time the diverse volcanic and sedimentary assemblages formed. Volcanic assemblages formed during this period include possible analogues to modern mafic plain, arc, rifted arc and back arc tectonic settings. Sedimentary assemblages types include both intragreenstone turbidite and alluvial-fluvial types, with minor alkalic metavolcanic rocks.

Early 3.17 to 2.83 billion-year-old calc-alkalic plutons are recorded within plutonic subprovinces and at margins of the central part of the Wabigoon Subprovince (Davis and Jackson 1988), but no metallic mineral deposits appear to have formed during this stage in this plutonic subprovince.

## METALLIC MINERAL DEPOSITS IN OLD GREENSTONE BELT ASSEMBLAGES (2.8 TO 3.03 Ga)

Known major metallic mineral deposits in the old (2.8 to 3.03 Ga) greenstone belt assemblages are limited to gold. Banded, oxide-facies iron formation units formed within shallow-water platform or platform-foredeep successions. Komatiite flows occur in platform assemblages, but deposits of komatiite-associated nickel-copper-platinum group element mineralization are not known (Thurston and Chivers 1990). No volcanic-associated, massive, base metal sulphide (VMS) deposits are known to occur in the old greenstone belt assemblages in the Superior Province of Ontario. Potential for detrital gold and/or uranium deposits in metasedimentary rocks in the platform sequences has been inferred, but not demonstrated (Thurston and Chivers 1990). Parts of the northwestern Superior Province may have experienced a collisional-type orogeny that terminated at about 2.9 Ga (Stott et al. 1989; Thurston et al., this volume). At least one gold deposit (Musselwhite), in the North Caribou greenstone belt (Breaks, Bartlett et al. 1985; Hall and Rigg 1986) may have formed during this orogeny.

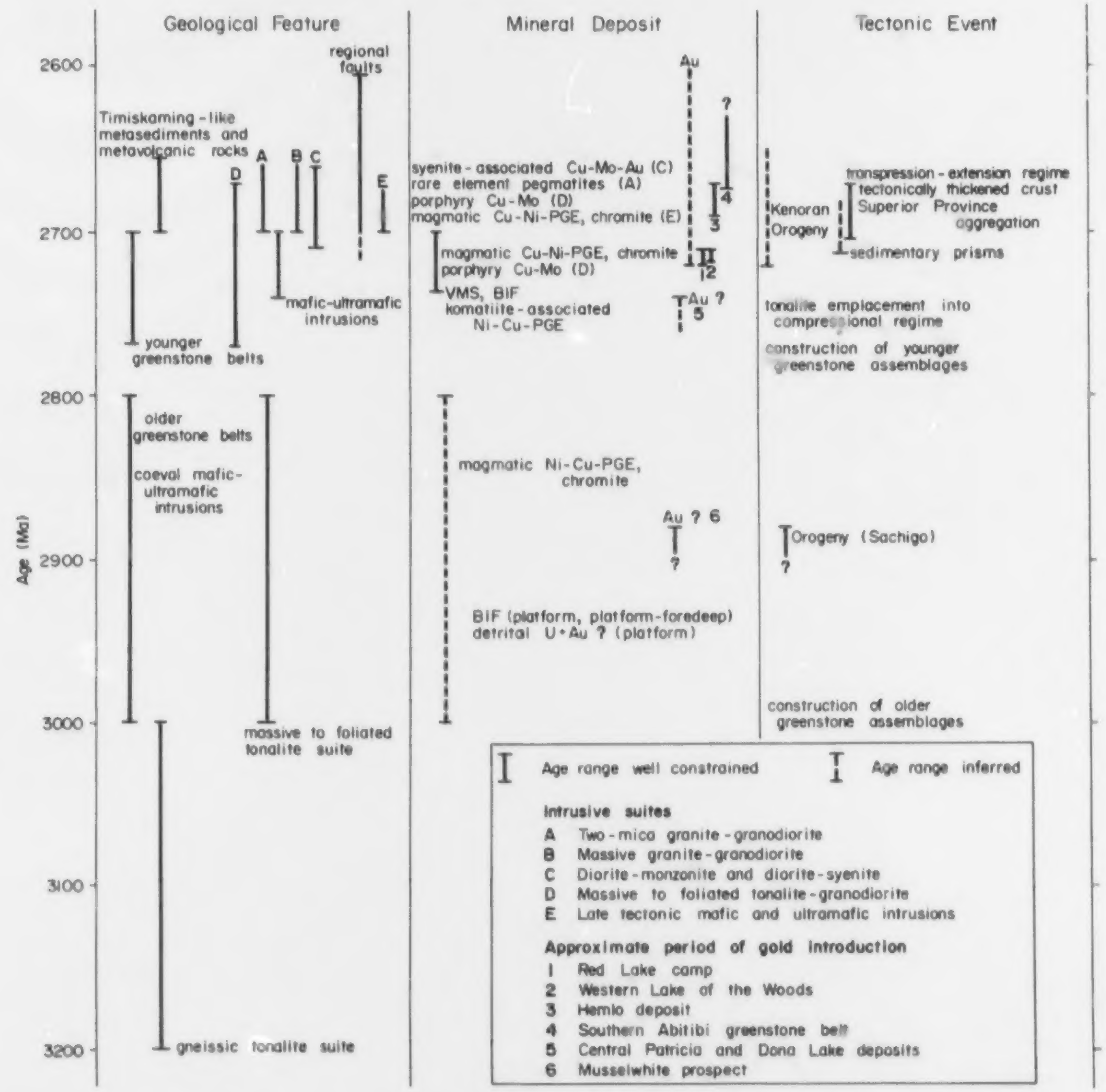

**Figure 22.22.** Synthesis of tectonic history of Superior Province and related metallic mineral deposits.

## METALLIC MINERAL DEPOSITS IN YOUNGER GREENSTONE BELT ASSEMBLAGES (2.7 TO 2.8 Ga)

Between 2.8 and 2.7 Ga, more diverse assemblages formed in disparate tectonic settings in the ancestral parts of the Uchi, Wabigoon, Wawa and Abitibi subprovinces. Metallic mineral deposits that formed contemporaneously with these assemblages and cogenetic intrusions include: 1) volcanic-associated, massive, base metal (VMS) deposits, 2) komatiite-associated nickel-copper-platinum group element deposits, 3) banded iron formation (BIF) deposits, 4) porphyry copper-molybdenum and copper-gold deposits, and 5) nickel-copper-platinum group element deposits associated with synvolcanic and postvolcanic mafic and ultramafic intrusions. Many gold deposits formed late in the tectonic history of the Superior Province and are superimposed upon pre-existing assemblages.

### VMS Deposits

Most VMS deposits formed between 2.75 and 2.7 Ga in successions that contain chemically evolved rhyolitic (high-silica rhyolite) and andesitic (icelandite) rocks. Parts of these volcanic successions, and the contained VMS deposits, were deposited during periods of extension in an arc, rifted arc, or back-arc setting. The Potter Mine VMS deposit occurs within a komatiite-tholeiite assemblage in the Abitibi Subprovince and may have formed in a rifted oceanic setting (Jackson and Fyon, this volume).

### Komatiite-Associated Nickel-Copper-Platinum Group Element Deposits

All known komatiite-associated nickel-copper-platinum group element deposits are younger than 2.8 billion years. The deposits occur within the Abitibi Subprovince and formed within extensional oceanic, rifted back-arc, or rifted arc settings. The Thierry deposit in the Uchi Subprovince and the Shebandowan deposit Wawa Subprovince are associated with deformed intrusions, which may have a komatiitic affinity.

### Banded Iron Formation (BIF) Deposits

Iron formation units accumulated between 2.8 and 2.7 Ga within, and adjacent to, volcano-plutonic arcs in greenstone belts. Iron formation associated with accretionary sedimentary prisms (e.g., metasedimentary subprovince) formed following the construction of the volcano-plutonic arcs, during a period of greenstone belt uplift and erosion. This event took place at about 2.7 Ga (Williams, Stott, Thurston et al., this volume). The least well described iron formation is that for which fragmentary evidence occurs in the Timiskaming and Timiskaming-like assemblages. These assemblages formed following the construction and assembly of the greenstone belts, early during the Kenoran Orogeny (*circa* 2.7 Ga), perhaps in an extensional environment.

### Porphyry Deposits

Few porphyry copper-molybdenum deposits occur in the Superior Province. Some formed in subvolcanic, foliated to massive granodiorite that ranges in age from less than 2.8 to 2.7 Ga (e.g., Beidelman Bay). These intrusions were emplaced during the calc-alkalic constructional phase of volcano-plutonic arcs. At least some of these intrusions formed in a collisional tectonic setting. Other trondhjemite intrusions, having chemical signatures suggestive of a tholeiitic parentage, may have been emplaced into an extensional setting.

### Mafic to Ultramafic Intrusion-Associated Nickel-Copper-Platinum Group Element and Chromite Deposits

Nickel-copper-platinum group element and chromite mineralization occurs within subvolcanic mafic to ultramafic komatiitic and tholeiitic intrusions that were emplaced during the construction of volcano-plutonic arcs and oceanic-like crust in greenstone belts. Generally these intrusions occur within komatiitic-tholeiitic assemblages. The intrusions may represent high-level magma chambers formed in an extensional oceanic setting or a back-arc basin, or during the primitive stages of arc construction.

## Deposits Formed During and Following the Assembly of the Greenstone Belts

There is mounting evidence that greenstone belts formed by the assembly of both autochthonous and allochthonous assemblages (e.g., Blackburn et al., this volume; Jackson and Fyon, this volume; Stott and Corfu, this volume; Thurston et al., this volume; Williams, Stott, Thurston et al., this volume). During, and following, the assembly of the disparate assemblages into volcanic-plutonic subprovinces, some gold deposition in the Uchi Subprovince and a second stage of porphyry copper-molybdenum deposits formed.

Some gold deposits in the central part of the Uchi greenstone belt (e.g., Dona Lake and Pickle Crow deposit) may have formed during the construction of subduction-related volcano-plutonic arcs, or during the assembly and juxtaposition of allochthonous assemblages, but prior to the regional folding and faulting related to the accretion of the Superior Province during the Kenoran Orogeny. Recognition of additional gold deposits that formed during construction, or assembly of volcano-plutonic arcs has been hindered by the superimposed tectonic effects of the Kenoran Orogeny.

The McIntyre porphyry copper-molybdenum deposit, Abitibi Subprovince, occurs within a 2.69 billion-year-old, postvolcanic granodiorite. This intrusion was emplaced 10 to 15 million years after the construction of the calc-alkalic Tisdale and Krist assemblages.

## Deposits Formed During the Erosion of Emergent Volcanic Material

During the final stages of arc coalescence, the granite-greenstone subprovinces were uplifted due to tectonic thickening. The resultant erosion of emergent volcanic terranes yielded detritus which filled prismatic sedimentary basins. The final stages of arc coalescence and deposition of sediment in accretionary prisms took place between 2.7 and 2.69 Ga (Williams, Stott, Thurston et al., this volume). No metallic mineral deposits appear to have formed during this regional-scale emergence.

## Progressive Crustal-Scale Shortening and Accretion of the Superior Province

The Uchi, English River, Winnipeg River, Wabigoon, Quetico, Wawa, and Abitibi subprovinces were involved in a southward younging, accretionary collisional event between 2720 and 2660 Ma, termed the Kenoran Orogeny (see Williams, Stott, Thurston et al., this volume). During this event, the Superior Province formed by accretion of granite-greenstone subprovinces and adjacent sedimentary prisms (Beakhouse 1986; Stott et al. 1987; Corfu 1988; Davis and Smith 1989; Williams, Stott, Thurston et al., this volume). A vertical component of slip is recorded along regional deformation zones, close to, or along, subprovince boundaries. In response to underthrusting, tectonic thickening in the sedimentary prism at subprovince margins resulted. Metamorphism of the tectonically thickened crust resulted in localized partial melting (e.g., Breaks, this volume; Williams, Stott, Thurston et al., this volume) to yield peraluminous and calc-alkalic tonalitic magmas. These magmas were emplaced into the sedimentary prisms. The peraluminous magmas crystallized to yield two-mica granites. Some of these two-mica granites underwent extensive chemical fractionation to yield rare element-bearing pegmatites (Breaks, this volume). The parental S-type granites and derivative rare element-bearing pegmatites were emplaced along subprovince boundaries, adjustment structures, and less commonly, within subprovinces. Some late mafic intrusions, emplaced along or near regional deformation zones, are host to copper nickel-platinum group element mineralization.

Few porphyry copper-gold deposits are documented in the Superior Province. Those in the Matachewan area, Abitibi greenstone belt, are spatially associated with the Larder Lake–Cadillac regional deformation zone, alkalic intrusions, and metavolcanic and metasedimentary rocks of the Timiskaming assemblage. The *circa* 2.68 billion-year-old alkalic intrusions are likely cogenetic with the alkalic metavolcanic rocks in the Timiskaming assemblage.

The tectonic setting in which the alkalic igneous and sedimentary rocks of the Timiskaming assemblage developed remains debated. The preserved distribution of Timiskaming and Timiskaming-like assemblages is consistent with deposition in pull-apart basins during transpression (Colvine et al. 1988; Card et al. 1989; Thurston and Chivers 1990). However, remnants of Timiskaming-like rocks and likely cogenetic alkalic intrusions are areally more widespread and not exclusively preserved adjacent to major faults. It is possible, therefore, that Timiskaming strata were initially more widespread and that their accumulation was controlled by volcanic landscape and possibly local faults (Corfu et al. 1991). The geochemistry of the Timiskaming alkalic igneous rocks is similar to that of alkalic rocks in the Roman Province, Italy (Cooke and Moorhouse 1969; Capdivila et al. 1982; Ujike 1985). A preferred tectonic setting for the Roman suite is that of a mature arc (Peccerillo 1985). Therefore, alkalic magmatism of the Timiskaming and Timiskaming-like assemblages may have been generated during the late stages of subduction during, and following, collision and amalgamation of arcs, or arc-continent collision (Corfu et al. 1991). The alkalic intrusion-associated copper-gold deposits and occurrences in the Abitibi and Shebandowan greenstone belts appear to have formed in this tectonic setting.

## Lateral, Subprovince-Parallel Slip

Between 2.7 and 2.66 Ga, continued shortening across the Superior Province was accommodated, in part, by a combination of under-thrusting and lateral, subprovince-parallel slip. Gold deposits formed late during, or following this tectonic stage, in and adjacent to regional deformation zones. These gold deposits and host structures were superimposed upon pre-existing crustal and metallogenic architecture and a direct relationship between gold and a specific magmatic type remains equivocal.

## Summary and Questions

The data available indicate that the formation of similar assemblage types was repeated in time and space through the Archean record. This illustrates that similar tectonic processes were operative at different times. Similarly, the juxtaposition of coeval assemblages that differ in rock association suggests that the juxtaposition of assemblages represents different, but contemporaneous, tectonic environments. Thus, komatiite-associated nickel-copper-platinum group element deposits formed at about the same time as, but in a different tectonic setting from some VMS deposits. Similarly, some gold deposits in the Uchi Subprovince may have formed during the construction, or juxtaposition, of volcan[illegible]onic arcs, when elsewhere, VMS deposits were about t[illegible]m.

Most metallic mineral resources occur in the younger (less than 2.8 Ga) greenstone belts. This is not typical of "older" greenstone belts preserved in other Archean cratons (Groves and Batt 1984). Further study of these "older" greenstone belts is required to carefully document their assemblage character, tectonic history, and characteristics and types of mineral occurrences.

A significant development in the gold metallogeny is the evidence in support of pre–Kenoran deposit formation. The inference that some gold deposits in the Uchi

Subprovince formed during the construction and/or accretion of volcano-plutonic arcs warrants additional investigation to further test this hypothesis, document other possible settings where such deposits may occur, and to re-examine known gold deposits, in areas where a Kenoran orogenic record is not prominent, to determine if any a syn-constructional age is consistent with observed geological relationships. The possibility that gold deposits in the Sachigo Subprovince formed during an orogeny at 2.86 Ga also warrants further testing. A spectrum of gold occurrences in the Sachigo Subprovince, which occur in disparate settings, must be studied and constrained geochronologically to test the existence of these "old" gold deposits.

Given the strong magmatic association inferred for gold, copper-gold, and copper-molybdenum deposits in Miocene and Recent arc systems (Sillitoe 1989), the apparent paucity of these magmatic deposits in the Superior Province of Ontario is anomalous. Few unequivocal porphyry copper deposits are documented; yet, apparently arcs and comagmatic intrusions exist. Similarly, are the gold-copper deposits in the Matachewan area of the Abitibi greenstone belt analogous to modern alkalic-associated copper-gold deposits?

Future investigations will continue to document the empirical characteristics of known deposits, their geological settings, and the relationships between the deposits and tectonic elements. The objective is to test more fully the framework summarized here and to enhance our understanding and our predictive ability to discover new metallic mineral resources.

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# Chapter 23

# Metallogeny of the Proterozoic Eon, Northern Great Lakes Region, Ontario

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## Abstract

**Paleoproterozoic and Mesoproterozoic rocks in the northern Great Lakes region of Ontario evolved over a period of 1.5 billion years, during several distinct tectonic superepisodes. These tectonic superepisodes, defined by the clustering of tectonic events in time that culminated in an orogeny, are separated by an apparent absence of tectonic activity. During this protracted tectonic history, a spectrum of mineral deposit types formed in diverse tectonic settings, during discrete tectonic superepisodes. Three distinct Proterozoic tectonic superepisodes, each temporally associated with the formation of metallic mineral deposits, have been documented in the Lake Superior region: 1) deposition of the Paleoproterozoic Animikie continental margin sedimentary prism, between 2.1 and 1.85 Ga, prior to, and during, the 1.86 billion-year-old Penokean Orogeny and emplacement of carbonatite and alkalic complexes at about 1.9 Ga; 2) intrusion of circa 1.54 billion-year-old, Mesoproterozoic anorogenic granites and deposition of the rift-related Sibley Group; and 3) Mesoproterozoic rifting along the Midcontinent Rift System between circa 1.11 and 1.09 billion years, during which the Keweenawan Supergroup was deposited. Metallic mineral deposits formed during the earliest, Paleoproterozoic event in the Lake Superior region include the deposition of iron-rich sedimentary rock in the rifted-passive margin and synorogenic foredeep sequences of the Animikie Basin and the formation of volcanic-associated, massive, base metal sulphide mineralization in volcanic arc successions of the Wisconsin Magmatic Terrane. Between 1.95 Ga and 1.80 Ga, in the Lake Superior area, collision along the Trans-Hudson Orogen resulted in the suturing of several Archean terranes. No known metallic mineral deposits formed in Ontario during this event, although komatiite-associated nickel-copper-platinum group element and epigenetic gold mineralization formed elsewhere in the Trans-Hudson Orogen. The Mesoproterozoic, 1.53 billion-year-old anorogenic granites of the Lake Nipigon area have potential to host yttrium, zirconium, rare earth element and tin mineralization, although deposits are not known. Clastic sedimentary sequences that were deposited in the rift-related, 1.5 to 1.3 billion-year-old Sibley Group, are host to unconformity-related uranium mineralization and red-bed-type copper mineralization. During the Mesoproterozoic Keweenawan rifting event, the most diverse spectrum of deposit types formed in the Lake Superior area: 1) magmatic nickel-copper-platinum group element and chromite mineralization formed in Keweenawan gabbro intrusions and the Port Coldwell alkalic complex; 2) Keweenawan basalt flows are host to disseminated native copper that fills amygdules and bornite-chalcocite-quartz-calcite veins; 3) amethyst and sulpharsenide silver-cobalt veins formed in, and adjacent to, faults that cut Keweenawan lava flows; 4) copper sulphide mineralization was deposited in porphyry intrusion-related breccia pipes; and 5) uranium mineralization was concentrated along unconformities. The Oronto Group clastic sedimentary sequence, deposited late during the Keweenawan rift event, in response to continued subsidence following volcanism, is host to red-bed-type copper deposits.**

**The Paleo- and Mesoproterozoic history of the area east of Lake Huron is more complex than that documented in the Lake Superior region. At least 5 distinct tectonic superepisodes are documented. The first began with rifting of the composite Archean craton at about 2.45 Ga, along a series of east-striking structures. This rifting may have triggered the emplacement of 2.48 billion-year-old gabbro-anorthosite intrusions that are host to magmatic nickel-copper-platinum group element mineralization. Rift development may have also triggered eruption of the 2.45 billion-year-old metavolcanic rocks in the lower part of the Elliot Lake Group that are host to virtually all volcanic-associated massive base metal sulphide concentrations outside of the Sudbury Igneous Complex. Continued extension and attendant erosion resulted in deposition of a northward-onlapping, shallow-water, Huronian Supergroup clastic prism, upon a south-facing continental margin. Paleoplacer uranium deposits at Elliot Lake accumulated within quartz-pebble conglomerate units in the Matinenda Formation, within the lower part of the Huronian Supergroup. The Lorrain and lower Gordon Lake formations, part of which may have been deposited in a tropical environment, marked the end of this sedimentary period and are host to red-bed copper mineralization. Shale-hosted copper mineralization occurs within the Gowganda Formation. Between 2.39 and 2.33 Ga, the Murray and Creighton granitic plutons were intruded into lower Huronian sediments near Sudbury. The tectonic significance of these intrusions is not clear.**

**The second tectonic event that affected the area east of Lake Huron is marked by the deformation of the Huronian Supergroup and the subsequent intrusion of 2.22 billion-year-old Nipissing gabbro. Magmatic nickel-copper-platinum group element mineralization occurs within the Nipissing gabbro and polymetallic skarn mineralization developed where calcareous sediments of the Espanola Formation were intruded by Nipissing gabbro.**

**The Penokean Orogeny in the area east of Lake Huron is manifested by the northward overriding of the Huronian Supergroup clastic prism by an unknown, allochthonous terrane, crustal thickening,**

metamorphism and thrust faulting. The emplacement of the 1.85 billion-year-old norite of the Sudbury Igneous Complex, host to magmatic nickel-copper-platinum group element mineralization, was coeval with the Penokean Orogeny. This magmatism was possibly triggered by a meteorite impact, an endogenic process, or a combination of both. Broadly coeval with the Sudbury Igneous Complex and the Penokean Orogeny was the formation of sedimentary-associated, massive, base metal sulphide deposits that formed at the base of the Whitewater Group, within the Sudbury basin. These sedimentary rocks may be time equivalent to the foredeep Animikie sedimentary rocks of the Lake Superior area. Several types of polymetallic hydrothermal veins occupy Penokean structures, indicating that the veins formed during, or after, the Penokean orogeny, at least 0.3 billion years after the intrusion of the Nipissing gabbro.

Following the Penokean Orogeny, a period of 1.74 billion-year-old anorogenic magmatism in the Killarney Magmatic Belt may reflect the extensional phase of a circa 1.7 billion-year-old collisional orogeny that affected the northwestern part of the Grenville Province in Ontario. No metallic mineralization is known to be spatially associated with these intrusions. However, gold and copper-gold vein occurrences and deposits, and attendant 1.7 billion-year-old sodium metasomatism, cut rocks of the Huronian Supergroup and Sudbury Igneous Complex. This mineralization and attendant sodium metasomatism may have developed late during the circa 1.7 billion-year-old orogenic event.

A period of 1.45 billion-year-old anorogenic magmatism within the Killarney Magmatic Belt is coeval with 1.55 to 1.4 billion-year-old anorogenic magmatism in the Lake Nipigon and Wisconsin areas. These intrusions in the Lake Nipigon and Wisconsin areas are associated with elevated concentrations of fluorine, yttrium, zirconium, rare earth elements, and tin and may be related to the Eastern Granite-Rhyolite Province. The magmatism may indicate pre-Grenville, crustal extension.

Related to tectonic and magmatic activity along the Mesoproterozoic Midcontinent Rift System, native copper and base metal sulphide veins formed in, and adjacent to, Keweenawan fault zones, and the Tribag breccia- and Jogran felsic intrusion-associated copper mineralization formed in the Sault Ste. Marie–Batchawana area.

The west-northwest-striking 1.24 billion-year-old Sudbury diabase dike swarm that cuts rocks of the Huronian Supergroup, is deformed and metamorphosed within the Grenville Front Tectonic Zone, demonstrating that orogenic events, related to the Grenville Orogeny, had been initiated by 1.24 Ga in this part of the Grenville Orogen. The 1.2 to 1.0 billion-year-old Grenville Orogeny, which occurred in the area southeast the Huronian Supergroup, involved northwest-directed crustal-scale thrusting, along southeast-dipping structures and extensive reworking of 1.8 to 1.35 billion-year-old crustal material. No significant metallic mineral deposits appear to have formed in the Huronian Supergroup during the Grenville Orogeny.

## INTRODUCTION

In this chapter, the principal metallic mineral deposit types present in the Proterozoic rocks in the north and east of the Great Lakes region of Ontario, in the Penokean Orogen, are described. The time and space relationships between the metallic mineral deposits and the tectonic histories of these areas are synthesized into a metallogenic history. The metallogeny of the Grenville Province is discussed in Chapter 24.

The Archean Superior Province is bounded by the following 2.0 to 1.8 billion-year-old Proterozoic collisional and accretionary orogens (Figure 23.1; Hoffman 1989): 1) Trans-Hudson, including the Hudson Bay and northern Quebec segments; 2) Penokean; 3) New Quebec, formerly Labrador Trough; and 4) Grenville. A variety of tectonic settings are represented in these Proterozoic orogens. For example, outward from the Archean craton, the Penokean Orogen consists of a deformed, older passive-margin sedimentary prism (Huronian Supergroup); a synorogenic foredeep sedimentary prism (Animikie Group); and an allochthonous island arc complex (Wisconsin Magmatic Terrane). The Trans-Hudson Orogen consists of (Galley et al. 1990): the Churchill–Superior foreland belt boundary zone; an internal zone of juvenile Proterozoic crust; the Andean-type, Wathaman–Chipewyan magmatic arc batholith; and the Hearne Province hinterland belt. The bordering Archean terranes were variably deformed during the Proterozoic events and have regions of basement reactivation, thrusting and transcurrent shearing (Hoffman 1989). Characteristics of the Grenville Province are discussed in Chapter 19.

## TRANS-HUDSON OROGEN

The collisional Trans-Hudson Orogen (Hoffman 1988) separates the Archean Superior Province to the east from the Archean Wyoming and Hearne provinces to the southwest and northwest respectively (*see* Figure 23.1). In Ontario, juvenile Proterozoic crust, including possible relics of island arcs and oceanic crust, is preserved in the Hudson Bay segment of the Trans-Hudson Orogen (Hoffman 1989). The segment is exposed on the coast and islands of eastern Hudson Bay and on inliers on the southern mainland including the Sutton Inlier in Ontario. Parts of the Hudson Bay segment of the Trans-Hudson Orogen are covered by the Paleozoic rocks in the Hudson Bay Basin (Hoffman 1989). The Hudson Bay segment consists of the autochthonous Nastapoka homocline, the Belcher thrust and fold belt, interarc sediments of the Winisk trough, and the central Hudson Bay hinterland (Roksandic 1987). Between 1.91 and 1.79 Ga, accretion of island arcs and sediments resulted from north and northwestward subduction along the Trans-Hudson Orogen (Hoffman 1989).

Important gold, copper-zinc and nickel-copper-sulphide deposits occur in the Trans-Hudson Orogen. In Manitoba and Saskatchewan, nickel-copper-platinum group element and gold mineralization occurs in 2 different lithotectonic domains within the Trans-Hudson Orogen. In the Flin Flon belt, deposits of volcanic-associated, copper-zinc massive sulphide (VMS) and epigenetic, mesothermal gold occur within a deformed juvenile volcanic arc (cf., Galley et al. 1990). The VMS deposits occur within the 1.886 billion-year-old (Gordon et al. 1990) volcanic Amisk Group (Galley et al. 1990). The gold mineralization is associated with shear zones and related structures that postdate the 1.832 billion-year-old (Gordon et al. 1990) volcanic and sedimentary Missi Group (Galley et al. 1990). In the Thompson Nickel Belt, nickel-copper-platinum group element (PGE) mineralization occurs within serpentinized ultramafic sills within a shelf-foredeep foreland succession (Galley et al. 1990). This succession overlies the Archean Superior Province craton (cf., Galley et al. 1990).

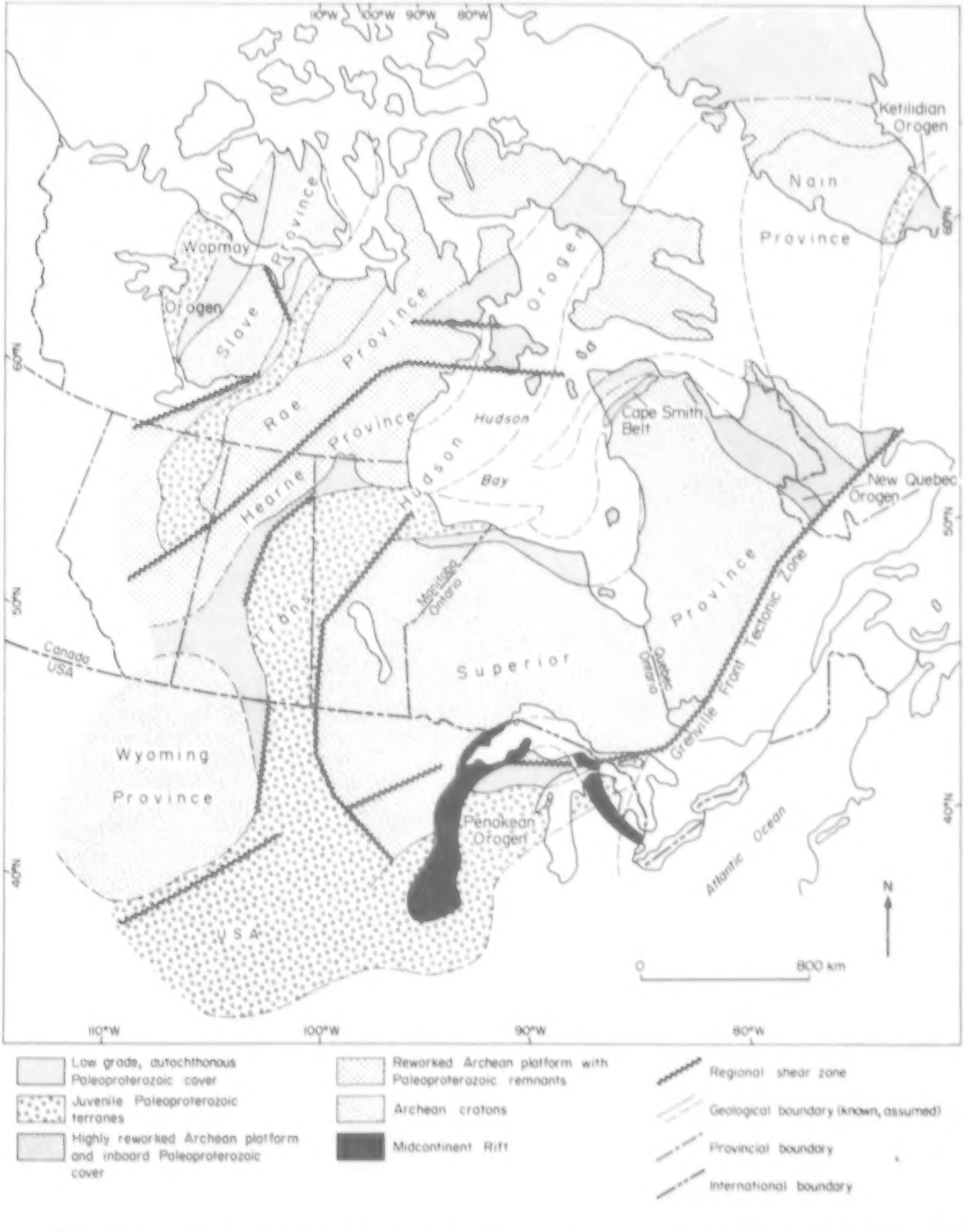

110°W
100°W
90°W
80°W
Ketilidian Orogen
Nain Province
60°N
Wopmay Orogen
Slave Province
Rae Province
Hearne Province
Orogen
Hudson
Hudson Bay
Cape Smith Belt
New Quebec Orogen
50°N
Trans-Hudson
Manitoba Ontario
Superior Province
Quebec Ontario
Grenville Front Tectonic Zone
Canada USA
Wyoming Province
Penokean Orogen
Atlantic Ocean
40°N
N
0
800 km
U.S.A.
Low grade, autochthonous Paleoproterozoic cover
Juvenile Paleoproterozoic terranes
Highly reworked Archean platform and inboard Paleoproterozoic cover
Reworked Archean platform with Paleoproterozoic remnants
Archean cratons
Midcontinent Rift
Regional shear zone
Geological boundary (known, assumed)
Provincial boundary
International boundary

**Figure 23.1.** Precambrian tectonic elements of northeastern North America. *Modified from* Hoffman 1989 and Galley et al. 1990.

In South Dakota, the Homestake Mining Company iron formation-hosted gold deposit occurs within a deformed succession of metavolcanic and metasedimentary rocks (cf., Slaughter 1968; Rye and Rye 1974). In the Cape Smith belt, in northern Quebec, nickel-copper-PGE sulphide deposits also occur in differentiated mafic and ultramafic sills and intrusions. These sills and intrusions formed feeder systems to komatiitic flows within a rift sequence (Giovenazzo et al. 1989). No metallic mineral deposits are yet known to occur in the Trans-Hudson Orogen in Ontario.

## PENOKEAN OROGEN

Flanking the Superior Province to the south is the Paleoproterozoic, Penokean Orogen (*see* Figure 23.1). Rocks in this orogen accumulated during rapid continental growth, between 1.80 and 1.65 Ga, following the suturing of Archean microcontinents between 1.95 and 1.80 Ga (Hoffman 1989). In the Lake Superior area (Figure 23.2), the Penokean Orogen includes the Animikie Group and Marquette Range Supergroup passive-margin sequence, and the oceanic Wisconsin Magmatic Terrane (*see* Sutcliffe, this volume). East of Lake Superior, the Penokean Fold Belt includes the southern part of the older Huronian Supergroup passive-margin prism succession that accumulated prior to, but was deformed during, the Penokean Orogeny (*see* Bennett et al., this volume).

The Penokean Orogen, in the western and southern Lake Superior areas, resulted from collision of island arcs of the Wisconsin Magmatic Terrane with the Archean craton and associated miogeoclinal sediments of the Marquette Range Supergroup and Animikie Group. This collision occurred along a southward-dipping subduction zone (Cambray 1978; Larue 1983). The 1.86 to 1.835 billion-year-old Niagara Suture Zone (*see* Figure 23.2) is interpreted as the present-day expression of the collision boundary between the 2 terranes (Cambray 1978; Larue 1983; Barovich et al. 1989).

### Lake Superior Area—Animikie Basin

Exposed in the Lake Superior area (*see* Figure 23.2) are Paleoproterozoic, Mesoproterozoic and Neoproterozoic sedimentary and igneous rocks (*see* Sutcliffe, this volume). The Proterozoic rocks are assigned to the Paleoproterozoic Animikie Basin and the Mesoproterozoic Midcontinent Rift system (Franklin et al. 1982; *see* Sutcliffe, this volume). The Proterozoic rocks, along the north shore of Lake Superior, are only slightly deformed and have undergone very low-grade metamorphism (Franklin et al. 1982). Those rocks, south of the lake, were deformed and metamorphosed during the Penokean Orogeny, particularly in proximity to the Great Lakes Tectonic Zone (Sims et al. 1980; Franklin et al. 1982). The Great Lakes Tectonic Zone defines the southern limit of the Superior Province basement rocks and the northern limit of the Minnesota River Valley gneiss terrane. The Great Lakes Tectonic Zone was also a fundamental factor in the development of the Paleoproterozoic strata, with thicker accumulations to the south (Sims et al. 1980).

## PALEOPROTEROZOIC ROCKS

South and west of Lake Superior, a 2.1 to 1.85 billion-year-old (Morey 1983) continental margin, sedimentary prism was deposited, within the Animikie Basin, on Archean basement. This basin is divided into a north zone of transgressive shelf clastic and chemical sedimentary rocks, a central zone of island arc igneous rocks and a south zone of volcanic-plutonic rocks (*see* Figure 23.2).

The north zone of the Animikie Basin is subdivided into northwestern and southeastern segments by the Midcontinent Rift (Morey 1983; *see* Sutcliffe, this volume). The northwestern segment contains sedimentary rocks of the Mille Lacs Group and the Mesabi and Cuyuna iron ranges of Minnesota. The Duluth Gabbro separates the northwestern segment from the Gunflint iron range of Ontario (*see* Figure 23.2; *see* Sutcliffe, this volume). These sedimentary rocks constitute a transgressive shelf sequence of shale, wacke and abundant iron formation. The southeastern region includes sedimentary rocks of the passive-margin Marquette Range Supergroup and equivalents (Mille Lacs and Chocolay Groups). These are overstepped northward by the synorogenic foredeep Animikie Group (*see* Figure 23.2; Cambray 1978; Larve 1983; Hoffman 1989).

The central zone of the Animikie Basin (*see* Figure 23.2) consists of the 1.88 to 1.76 billion-year-old (Van Schmus 1980) Wisconsin Magmatic Terrane (Barovich et al. 1989) which includes mafic and felsic metavolcanic rocks in the Crandon area and the Quinnessic Group. The Niagara Suture Zone separates the Wisconsin Magmatic Terrane from the Marquette Range Supergroup to the north (*see* Figure 23.2; Hoffman 1989). The southern limit of Archean crust, contiguous with the Superior Province, is the Niagara Suture Zone (Hoffman 1989). The Wisconsin Magmatic Terrane is interpreted as an accreted, oceanic, island-arc, igneous belt in which copper-zinc VMS deposits occur within dacite-rhyolite successions (Cambray 1978; May and Schmidt 1982; Larue 1983).

The south zone of the Animikie Basin (*see* Figure 23.2) consists of the Central Wisconsin rhyolite-plutonic complex. Volcanic rocks are dominantly subaerial, felsic in composition and are preserved as roof pendants in the anorogenic Wolf River Batholith (Franklin et al. 1982).

## MESOPROTEROZOIC SEDIMENTARY AND INTRUSIVE ROCKS OF THE NIPIGON EMBAYMENT

Mesoproterozoic, pre-Keweenawan sedimentary rocks of the Sibley Group, Keweenawan Nipigon sills, pre-Keweenawan alkali granite and related volcanic rocks and minor lamprophyre dikes occupy the Nipigon Embayment (*see* Figure 23.2; *see* Sutcliffe, this volume). The Sibley Group consists of quartz arenite, conglomerate, stromatolitic dolomite, shale and mudstones and is interpreted as a red-bed sequence (Coates 1972; Franklin et al. 1982; *see* Sutcliffe, this volume). The Sibley Group is preserved in a fault-bounded basin, south of Lake Nipigon (Coates 1972; Franklin et al. 1980; *see* Sutcliffe, this volume). The thickest part of the Sibley Group occurs against the Black Sturgeon fault, that may be part of a north-trending, asymmetric graben (*see* Sutcliffe, this volume). Some of the alkali granite intrusions in the Nipigon area (cf., English Bay Granite, Davis and Sutcliffe 1985) are characterized by high zirconium, yttrium and rare earth element (REE) abundances, consistent with an anorogenic (A-type), crustal-derived magma origin (*see* Sutcliffe, this volume). These intrusions are chemically similar to the 1.4 billion-year-old anorogenic Wolf River Batholith, in Wisconsin (Anderson and Cullers 1978), and some intrusions in the 1.74 to 1.45 billion-year-old Killarney Granitic

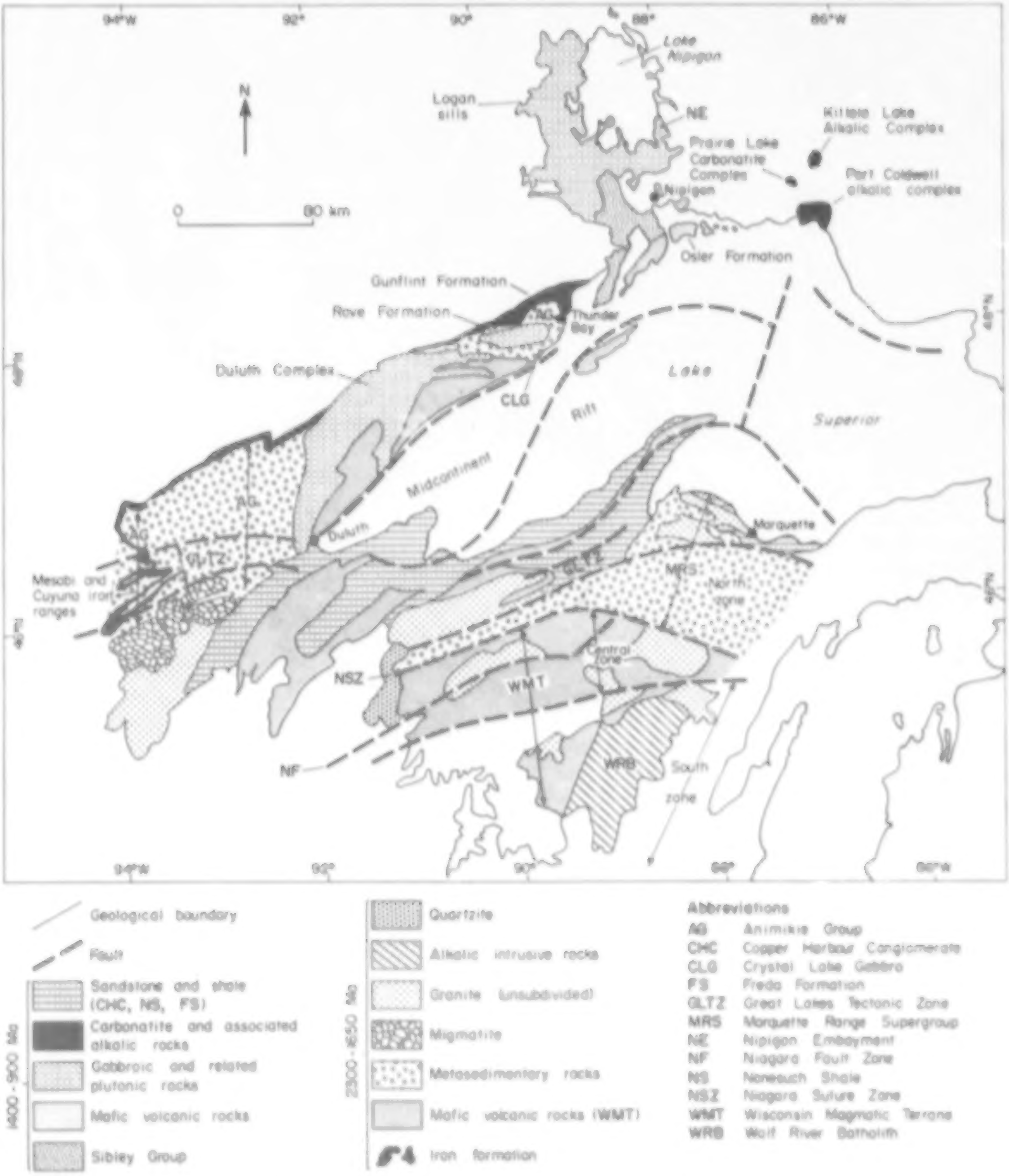

94°W
92°
90°
88°
86°W
48°N
47°N
46°N
N
0
80 km
Lake Nipigon
Logan sills
NE
Kitlola Lake Alkalic Complex
Prairie Lake Carbonatite Complex
Nipigon
Port Coldwell alkalic complex
Osler Formation
Gunflint Formation
Rove Formation
AG
Thunder Bay
Duluth Complex
CLG
Lake
Superior
Rift
Midcontinent
Duluth
AG
AG
GLTZ
Marquette
GLTZ
MRS
North zone
Mesabi and Cuyuna iron ranges
Central zone
NSZ
WMT
WRB
South zone
NF
Geological boundary
Fault
1400 - 900 Ma
Sandstone and shale (CHC, NS, FS)
Carbonatite and associated alkalic rocks
Gabbroic and related plutonic rocks
Mafic volcanic rocks
Sibley Group
2300 - 1850 Ma
Quartzite
Alkalic intrusive rocks
Granite (unsubdivided)
Migmatite
Metasedimentary rocks
Mafic volcanic rocks (WMT)
Iron formation
Abbreviations
AG Animikie Group
CHC Copper Harbour Conglomerate
CLG Crystal Lake Gabbro
FS Freda Formation
GLTZ Great Lakes Tectonic Zone
MRS Marquette Range Supergroup
NE Nipigon Embayment
NF Niagara Fault Zone
NS Nonesuch Shale
NSZ Niagara Suture Zone
WMT Wisconsin Magmatic Terrane
WRB Wolf River Batholith

**Figure 23.2.** Generalized distribution of Proterozoic rocks in the Penokean Orogen, Lake Superior area. *Modified from* Franklin et al. (1982) and Hoffman (1989).

Complex (*see* Easton, this volume). The Sibley Group occupies a north-trending block which resulted from a failed arm that developed about the Nipigon rift (Franklin et al. 1982).

### KEWEENAWAN SUPERGROUP

Tholeiitic, subaerial basalts, minor icelandite, rhyolite, minor felsic intrusions and minor clastic metasedimentary rocks of the Keweenawan Supergroup were deposited in, and marginal to, the Midcontinent Rift at approximately 1.1 Ga (*see* Figure 23.2; Van Schmus et al. 1982; *see* Sutcliffe, this volume). Lying stratigraphically above the Keweenawan volcanic successions are: 1) the Upper Keweenawan, red, shallow-water lacustrine sandstone and conglomerate of the Copper Harbour Conglomerate; 2) reduced, grey siltstone, shale and sandstone of the Nonesuch Shale; and 3) fluvial, red-bed sandstone of the Freda Formation. These successions represent a red-bed sequence that was deposited during regional subsidence, following the termination of volcanism (*see* Sutcliffe, this volume). This sequence is overlain by the Jacobsville Formation (Ojakangas and Morey 1982; *see* Sutcliffe, this volume).

Keweenawan intrusive rocks include: 1) tholeiitic, Logan and Nipigon dike swarms and sills, intruded at an early stage of rifting; 2) layered, tholeiitic intrusions such as the Duluth Complex in Minnesota and Crystal Lake Gabbro in Ontario; 3) mafic to felsic ring complexes; and 4) alkalic complexes such as the Port Coldwell alkalic complex (Weiblen 1982; Mitchell and Platt 1982; *see* Sutcliffe, this volume). The tholeiitic Duluth Gabbro, Crystal Lake Gabbro and the Pine River-Mount Mollie intrusion are associated with copper-nickel-PGE mineralization (Macdonald and Cherry 1988; *see* Sutcliffe, this volume). Mesoproterozoic carbonatite and alkalic complexes, such as the Port Coldwell and Killala Lake alkalic complexes and the Prairie Lake Carbonatite Complex, were emplaced along zones of regional faulting within the Superior Province (*see* Sage, this volume).

## Lake Huron Area—Huronian Supergroup

In the Sault Ste. Marie–Sudbury–Cobalt area, the southern margin of the Superior Province consists of the 1.85 billion-year-old (Krogh et al. 1984) Sudbury Igneous Complex and the 2.48 to 2.22 billion-year-old (Van Schmus 1976) Huronian Supergroup sedimentary prism and Cobalt Embayment (*see* Bennett et al., this volume). The Huronian Supergroup contains 4 unconformity-bounded, sedimentary prisms with minor amounts of basal volcanic rocks (cf., Frarey 1977; *see* Bennett et al., this volume; Sutcliffe, this volume). Thickness and facies variations reflect syndepositional, south-side-down normal faulting, which controlled the accumulation and preservation of the 3 lower prisms (Frarey 1977; Hoffman 1989; *see* Bennett et al., this volume). The 3 lower prisms are overlapped northward by the youngest prism which records postrift, regional subsidence (Zolnai et al. 1984). Several 2.48 billion-year-old (Krogh et al. 1984; McCrank et al. 1989) gabbro-anorthosite intrusions and related 2.45 Ga billion-year-old (Krogh et al. 1984) basalt-rhyolite volcanic successions, at the base of the oldest prism, are possibly indicative of initial rift development. The entire Huronian Supergroup is cut by 2.22 billion-year-old (Corfu and Andrews 1986) Nipissing gabbro intrusions.

The basal part of the Elliot Lake Group accumulated as a result of incipient rifting and contains the only recognized volcanic rocks in the Huronian Supergroup. In the western part of the area, the volcanic rocks are subaerial andesite and basalt of the Thessalon Formation (*see* Bennett et al., this volume). In the eastern or Spragge–Sudbury part of the area, the basal part of the Elliot Lake Group consists mainly of subaqueous tholeiitic metabasalt of the Pater, Salmay Lake, Elsie Mountain and Stobie formations and $2450^{+25}_{-10}$ million-year-old (Krogh et al. 1984) rhyolite of the Copper Cliff Formation (*see* Bennett et al., this volume). The upper part of the Elliot Lake Group consists of arkose, conglomerate, wacke, siltstone, dolomites and arenite (Card et al. 1984). The Elliot Lake Group is overlain by the Hough Lake, Quirke Lake and Cobalt groups consisting of conglomerate, wacke-siltstone and cross-bedded arenite units (Card et al. 1984). The Huronian Supergroup thickens southward across the east-striking Murray Fault System (Card et al. 1984). During the collisional Penokean Orogeny, the southern parts of the Huronian Supergroup were metamorphosed, folded and thrust northwards.

## Metallic Mineral Deposits

A spectrum of metallic mineral deposits occurs in the Proterozoic rocks of the Great Lakes region of Ontario. The mineral deposits are associated with both igneous and sedimentary rocks. The formation of some mineral deposits, such as the magmatic nickel-copper-PGE deposits, was coeval with their host rocks. Other deposit types, such as hydrothermal veins, postdate their host rocks and formed in response to later Proterozoic tectonic events.

In the Lake Superior area, the Paleoproterozoic successions contain relatively few metallic mineral deposit types, and there are currently no producing metallic mines in Ontario. For example, the south zone (*see* Figure 23.2) contains no metallic mineral deposits despite the presence of the A-type Wolf River Batholith. The central zone (*see* Figure 23.2), or Wisconsin Magmatic Terrane, contains several volcanic-associated, massive, base metal sulphide deposits, such as the Crandon deposit (May and Schmidt 1982; Rowe 1982). The northern zone (*see* Figure 23.2) contains the large Superior-type iron formations (James 1954). The Keweenawan rift successions in the Lake Superior area are associated with a more complex spectrum of metallic mineral deposit types, including (Franklin 1970; Franklin et al. 1980): 1) magmatic copper-nickel-PGE sulphide mineralization associated with mafic intrusions; 2) niobium-uranium mineralization associated with the Prairie Lake Carbonatite Complex; 3) hydrothermal copper, breccia-pipes; 4) base metal and silver-sulpharsenide hydrothermal veins; 5) uranium veins and disseminations associated with the unconformity between the Sibley Group and Archean rocks; and 6) red-bed copper mineralization.

In the Huronian Supergroup, deposits of paleoplacer uranium and magmatic nickel-copper-platinum group element (PGE) mineralization are currently being exploited. Gold values up to 600 times background are reported for sedimentary rocks of the Huronian Supergroup (Colvine 1981; Bennett 1983); however, more detailed studies (Tortosa 1984; Long and Colvine 1985; Rice 1987, 1988; Debicki 1990) have not identified paleoplacer gold deposits of commercial interest. Other metallic mineral deposit types present include (Innes and Colvine 1979, 1984; Pearson 1980; *see* Sutcliffe, this volume): 1) volcanic-associated, massive, base metal sulphide mineralization (VMS); 2) red-bed copper mineralization; 3) nickel-copper-PGE mineralization associated with mafic intrusions; 4) base metal-bearing skarn deposits; 5) hydrothermal base- and precious-metal veins; and 6) detrital, paleoplacer uranium deposits in the Elliot Lake area (*see* Bennett et al., this volume). Approximately 1000 base- and precious-metal vein occurrences are documented. Most are of minor importance; however, silver-sulpharsenide veins in the Cobalt silver area have yielded about 500 million ounces of silver since their discovery in 1903 (Owsiacki 1984). Other sites of historic metallic production include copper from the Batchawana area and Michipicoten Island and silver from the Thunder Bay area.

## IRON FORMATION

Iron ore has been produced from the Paleoproterozoic successions in the north zone of the Lake Superior area (*see* Figure 23.2); that is, in the Mesabi Iron Range within the Marquette Range Supergroup in Minnesota. Although magnetite taconite iron formation occurs in the Gunflint Formation in Ontario (James 1954; Goodwin 1956, 1960), there has been no commercial iron ore production.

## VOLCANIC-ASSOCIATED, MASSIVE, BASE METAL SULPHIDE MINERALIZATION

Base metal, sulphide-rich mineralization, having some characteristics akin to Phanerozoic VMS deposits (Pearson 1980), occurs within metavolcanic rocks in the lower part of the Huronian Supergroup, between Blind River and Sudbury (Figure 23.3; Pearson 1980). These deposits occur within (Pearson 1980): 1) mafic metavolcanic rocks of the Salmay Lake and Stobie formations; 2) cherts (Pearson 1980) or silicified volcanic rocks within successions of mafic metavolcanic rock of the Salmay Lake Formation; and 3) interflow metasedimentary units within mafic metavolcanic rock successions of the Stobie Formation. The Salmay Lake and Stobie formations and contained VMS mineralization is inferred to be 2.48 billion-years-old (Krogh et al. 1984; McCrank et al. 1989). The 2.45 billion-year-old (Krogh et al. 1984) Copper Cliff rhyolite caps the lower volcanic succession of the Huronian Supergroup.

The Pater Mine (*see* Figure 23.3) is an example of VMS mineralization that occurs within the basal Huronian metavolcanic rocks (Pearson 1980). This deposit, which occurs within the Pater Volcanics near Spragge, produced about 70 million pounds of copper and by-product gold and silver between 1960 and 1970 (Shklanka 1969; Robertson 1970; Pearson 1980). Mineralization consisted of pyrrhotite, chalcopyrite, pyrite and rare cobaltite (Knight 1963) and occurred in a thin sheet-like zone of brecciated country rock, of possible tectonic origin (Pearson 1980). The mineralization occurs along the southern margin of a silicified zone, in volcanic rocks, which extends along strike for at least 900 m (Knight 1963). Although interpreted as an exhalative chert (Pearson 1980), the silica-rich rock may also represent an alteration zone oriented parallel to the Murray Fault.

VMS copper mineralization in the Massey–Agnew Lake area (*see* Figure 23.3), including the Massey Mine and the Alexander, Bishop and Noranda Shakespeare prospects, occurs associated with silica-rich rocks (Pearson 1980). The silica-rich units are overlain by, interlayered with, or occur in close proximity to, deformed mafic metavolcanic rocks of the Salmay Lake Formation, close to metasedimentary rocks of the McKim and Matinenda formations (Pearson 1980). Mineralization consists of "net-textured", massive, blebby, stringer and disseminated chalcopyrite and pyrite (Pearson 1980). Contained metals are copper and iron, with minor cobalt and uranium (Pearson 1980). Precious metal tenors are very low (Pearson 1980). Mineralization at the Spanish River Mine is associated with elevated uranium concentrations (Pearson 1980). At the Massey Mine, several different types of mineralized and unmineralized breccia, including quartz-cemented breccia, are present. In proximity to these copper occurrences, Archean granitic rocks near the basal unconformity of the Huronian Supergroup are also silicified (Pearson 1980). It is uncertain if the silicification of the Archean granitoid rocks is temporally related to the silica replacement of the volcanic rock.

VMS mineralization also occurs within clastic metasedimentary units, which are interlayered with mafic metavolcanic rocks of the Stobie Formation, in the lower part of the Huronian Supergroup (Pearson 1980). Repetitive volcanic and sedimentary units consist of a lower basaltic flow unit, overlain by sulphide-bearing siltstone and wacke, followed by sulphide-poor pelite and arkosic arenite (Card et al. 1977). Stratabound sulphide mineralization is concentrated along, and just above, brecciated metavolcanic-metasedimentary contacts, principally within thin (less than 3 m) metapelites (Card et al. 1977; Pearson 1980). Pyrrhotite, chalcopyrite and pyrite occur disseminated and as stratiform layers within, and along fractures that cut the metapelites (Pearson 1980).

Within Proterozoic rocks of the Lake Superior area, younger, volcanic-associated, massive, base metal sulphide deposits (VMS), typified by the Crandon deposit (May and Schmidt 1982), occur within the 1.88 to 1.76 billion-year-old (Van Schmus 1980) central zone (*see* Figure 23.2), or Wisconsin Magmatic Terrane (Barovich et al. 1989). These deposits occur within successions of andesite and dacite pyroclastic rock (May and Schmidt 1982). The Wisconsin Magmatic Terrane and contained VMS deposits compose part of an allochthonous, accreted oceanic island-arc igneous belt (Cambray 1978; May and Schmidt

1982; Larue 1983). No known equivalent Proterozoic volcanic rocks or VMS deposits of comparable age occur in Ontario, although gneisses in the northwestern Central Gneiss Belt of the Grenville Province (*see* Easton, this volume) may be equivalent.

## SEDIMENTARY-ASSOCIATED, MASSIVE, POLYMETALLIC SULPHIDE MINERALIZATION

The Errington and Vermilion zinc-lead-copper-silver-gold sulphide deposits (*see* Figure 23.3) are stratabound, massive and disseminated polymetallic deposits that occur in the Whitewater Group, overlying the Sudbury Igneous Complex, in the Sudbury Basin. The deposits occur within the Vermilion Member approximately along the contact between the Onaping and Onwatin formations (Card and Hutchinson 1972; Sangster 1972; Pearson 1980; Rousell 1984). The Vermilion Member consists of dolomite, limestone, chert breccia and mudstone (Martin 1957). Disseminated and massive sulphide mineralization consists of pyrite, sphalerite, galena, chalcopyrite, marcasite and pyrrhotite (Rousell 1984). The disseminated mineralization occurs in the carbonate-rich rock, while the massive mineralization generally occurs at the base of the carbonate-rich rock and replaces the underlying mudstone (Rousell 1984). The base of the Whitewater Group is cut by 1.85 billion-year-old (Krogh et al. 1984) norite of the Sudbury Igneous Complex, providing a minimum age for the mineralization.

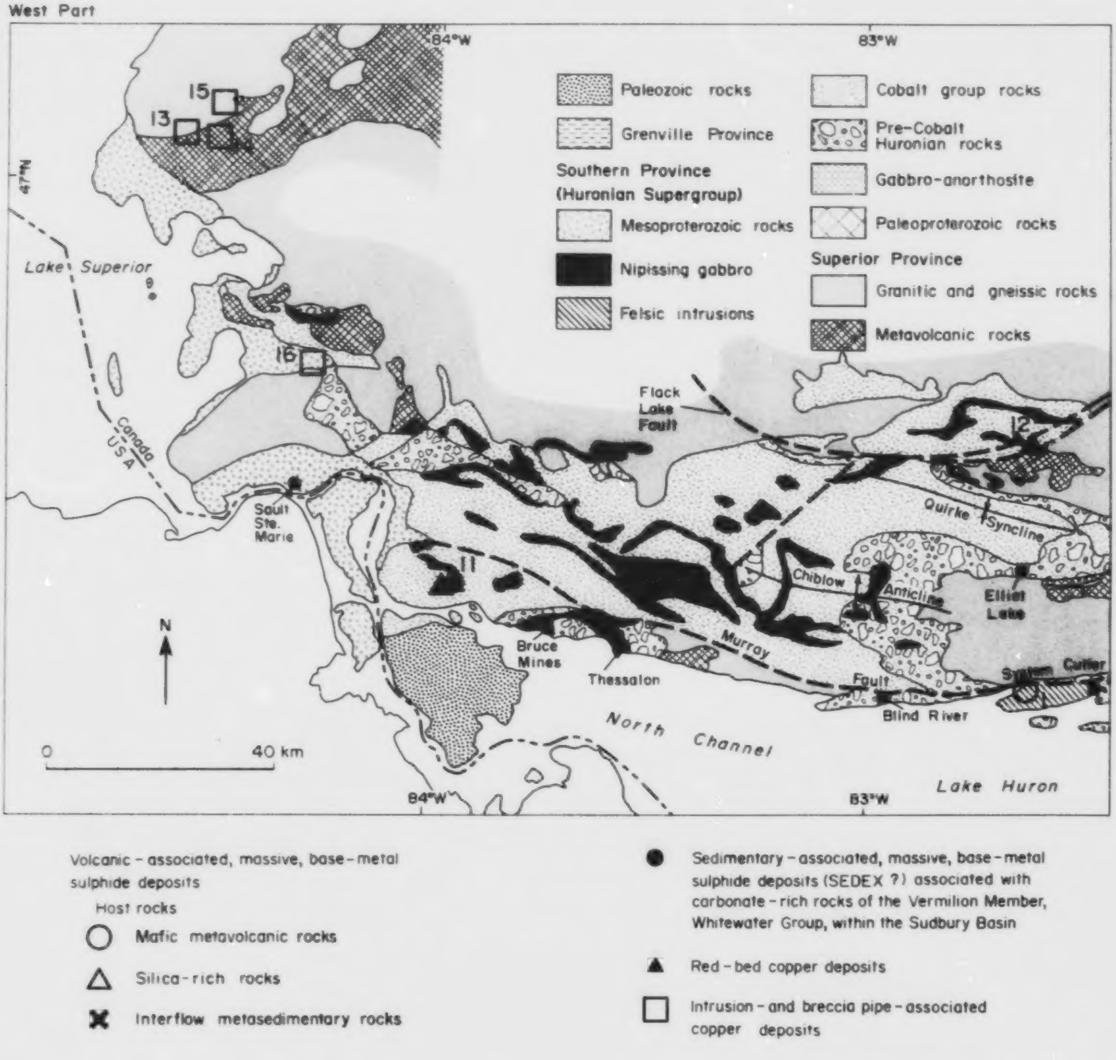

West Part
84°W
83°W
47°N
15
13
16
11
12
Paleozoic rocks
Grenville Province
Southern Province (Huronian Supergroup)
Mesoproterozoic rocks
Nipissing gabbro
Felsic intrusions
Cobalt group rocks
Pre-Cobalt Huronian rocks
Gabbro-anorthosite
Paleoproterozoic rocks
Superior Province
Granitic and gneissic rocks
Metavolcanic rocks
Lake Superior
8
Canada
USA
Sault Ste. Marie
Flack Lake Fault
Quirke Syncline
Chiblow Anticline
Elliot Lake
Murray Fault
Bruce Mines
Thessalon
Blind River
Cutler
North Channel
Lake Huron
N
0
40 km
Volcanic-associated, massive, base-metal sulphide deposits
Host rocks
Mafic metavolcanic rocks
Silica-rich rocks
Interflow metasedimentary rocks
Sedimentary-associated, massive, base-metal sulphide deposits (SEDEX ?) associated with carbonate-rich rocks of the Vermilion Member, Whitewater Group, within the Sudbury Basin
Red-bed copper deposits
Intrusion- and breccia pipe-associated copper deposits

**Figure 23.3.** Locations of Proterozoic, volcanic- (VMS) and sedimentary-associated (SEDEX) massive sulphide mineralization and sedimentary, red-bed-type copper deposits within Proterozoic rocks between Sudbury and Sault Ste. Marie (*modified from* Pearson 1980). VMS deposits—1 to 7; SEDEX deposits—8 to 10; red-bed-type copper deposits—11 and 12; and intrusion- and breccia pipe-associated copper deposits 13 to 16.

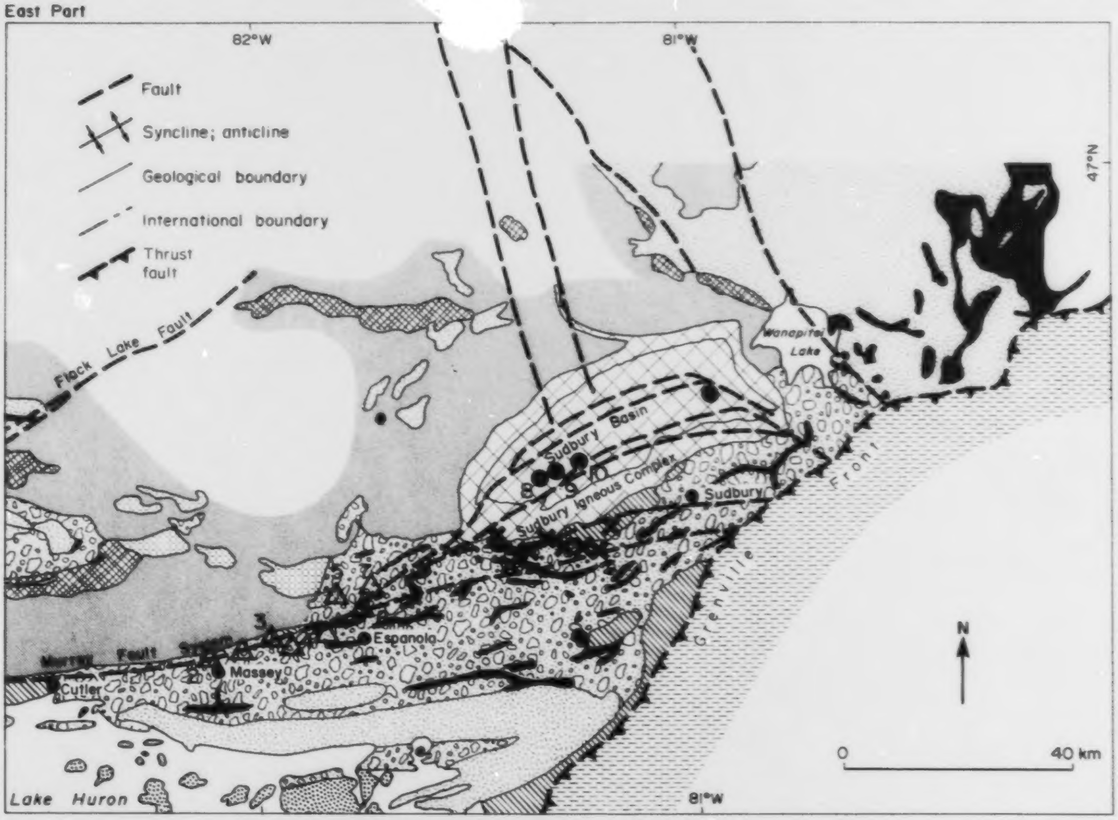
The origin of these deposits was initially attributed to replacement processes (Martin 1957); more recently, they have been reinterpreted as syngenetic, stratabound massive sulphide deposits (Card and Hutchinson 1972; Rousell 1984). The mineralization occurs within a rock succession that is similar to those of Phanerozoic sedimentary-associated, stratabound lead-zinc (SEDEX) deposits and may, therefore, have formed in a similar geological environment (Rousell 1984).

## NICKEL-COPPER-PLATINUM GROUP ELEMENT MINERALIZATION

Magmatic nickel-copper-platinum group element (PGE) mineralization formed during several mafic magmatic episodes: 1) 2.48 billion-year-old gabbro-anorthosite intrusions; 2) 2.22 billion-year-old Nipissing gabbro; 3) 1.85 billion-year-old norite and Sublayer rocks of the Sudbury Igneous Complex; 4) *circa* 1.0 billion-year-old Keweenawan gabbro-anorthosite intrusions; and 5) 1.1 billion-year-old Keweenawan alkalic complexes.

### Nickel-Copper-Platinum Group Element Mineralization Associated with 2.48 Billion-Year-Old Gabbro-Anorthosite Intrusions

A few nickel-copper-PGE occurrences (Figure 23.4) are associated with 2.48 billion-year-old (McCrank et al. 1989) East Bull Lake and Shakespeare–Dunlop gabbro-anorthosite intrusions. Mineralization in the East Bull Lake Gabbro-Anorthosite Intrusion consists of (Peck and James 1989): 1) disseminated to semimassive pyrrhotite and


East Part
82°W
81°W
Fault
Syncline; anticline
Geological boundary
International boundary
Thrust fault
Flack Lake Fault
Wanapitei Lake
Sudbury Basin
Sudbury Igneous Complex
Sudbury
Grenville Front
Espanola
Massey
Cutler
Murray Fault System
3
8
9
10
Lake Huron
47°N
N
0
40 km
81°W

1 Pater Mine
2 Massey Mine
3 Alexander
4 Bishop
5 Shakespeare
6 Springer
7 Spanish River Mine
8 Vermilion Mine
9 Sturdy
10 Errington Mine
11 Desbarats
12 Stag Lake and Cobre Lake area
13 Jogran
14 Palmer
15 Tribag
16 Nystedt

**Figure 23.3.** Locations of Proterozoic, volcanic- (VMS) and sedimentary-associated (SEDEX) massive sulphide mineralization and sedimentary, red-bed-type copper deposits within Proterozoic rocks between Sudbury and Sault Ste. Marie (*modified from* Pearson 1980). VMS deposits—1 to 7; SEDEX deposits—8 to 10; red-bed-type copper deposits—11 and 12; and intrusion- and breccia pipe-associated copper deposits 13 to 16.

chalcopyrite within, and adjacent to, east-striking shear zones in a variety of igneous host rocks; 2) pyrrhotite and chalcopyrite in nodular gabbroic anorthosite, and anorthosite, at the basal and outer part of the intrusion (Figure 23.5); and 3) pyrrhotite and chalcopyrite within an altered, marginal phase of the intrusion. The common localization of mineralized zones in marginal domains of the intrusion, characterized by the presence of a high proportion of wall rock xenoliths, is consistent with the assimilation of wall rock as the mechanism that triggered sulphur saturation (Pearson 1980, p.181). Although the East Bull Lake Gabbro-Anorthosite Intrusion is located approximately 25 km north of the Murray Fault, many nickel-copper-PGE occurrences are associated with smaller, elongate gabbro-anorthosite intrusions that are spatially associated with the Murray Fault (Giblin et al. 1979). The Murray Fault, therefore, appears to have localized the emplacement of some of these early gabbro-anorthosite intrusions.

## Nickel-Copper-Platinum Group Element Mineralization Associated with Nipissing Gabbro Intrusions

Nickel-copper-PGE sulphide concentrations associated with Nipissing gabbro occur principally in 2 clusters, although Nipissing gabbro is areally widespread (*see* Figure 23.4). One cluster occurs southwest of the Sudbury Igneous Complex, where intrusions of Nipissing gabbro cut rocks of the Elliot Lake and Hough Lake groups (*see* Figure 23.4). These occurrences lie south of the Murray

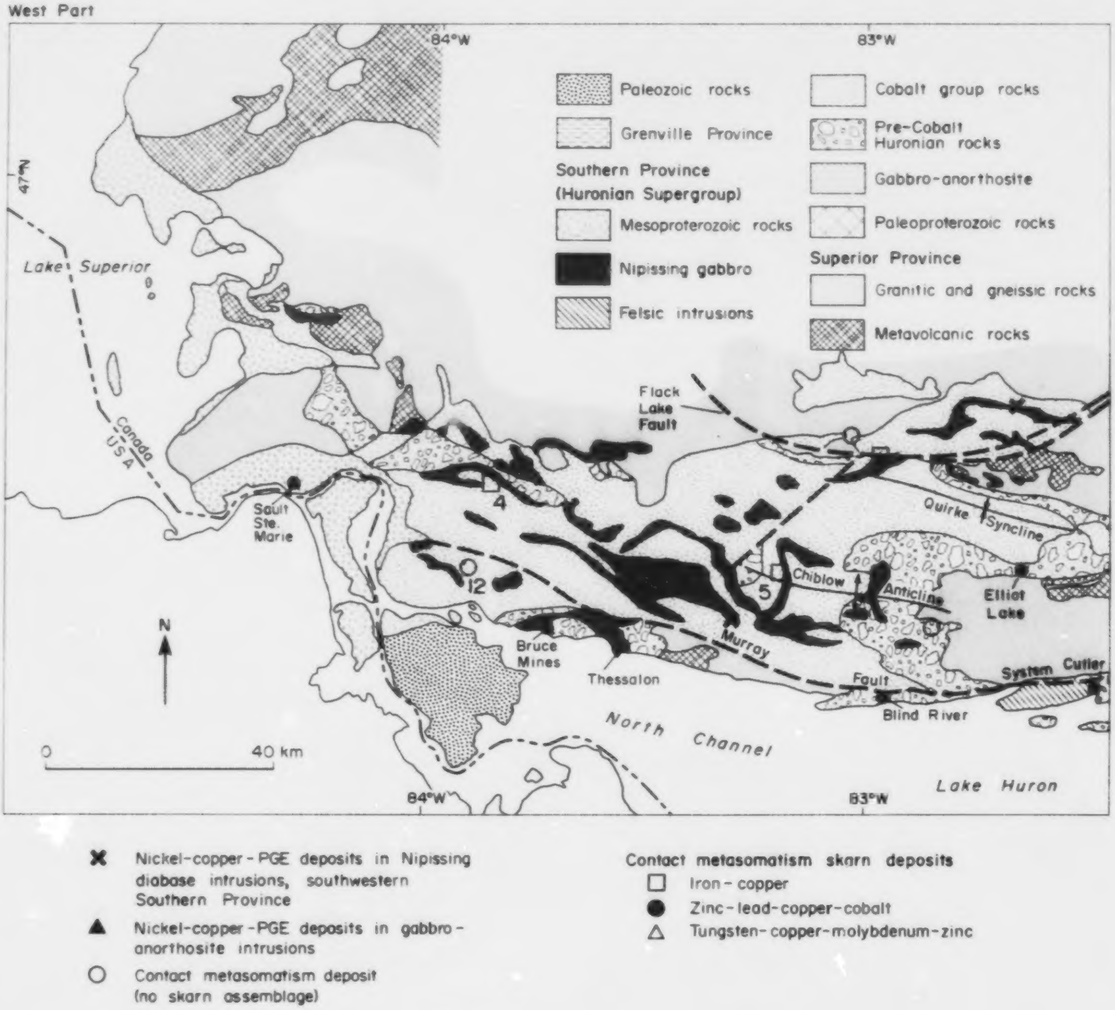
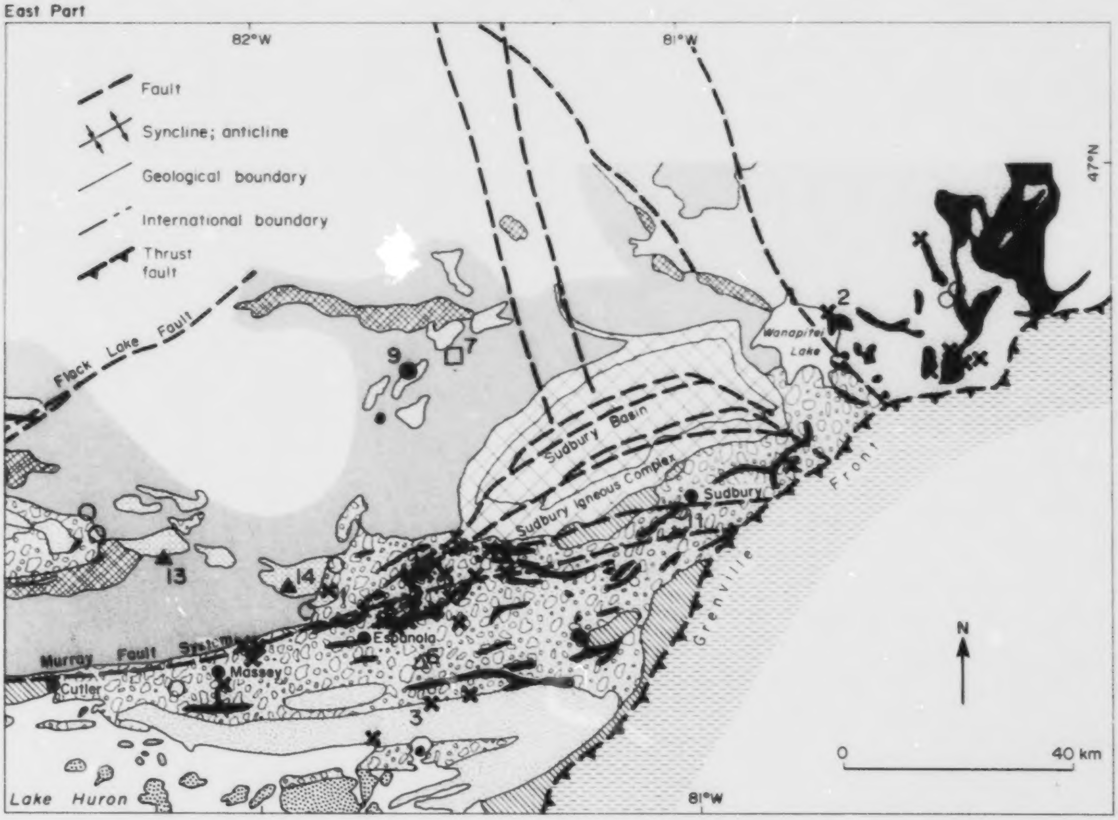

West Part
84°W
83°W
47°N
Paleozoic rocks
Grenville Province
Southern Province
(Huronian Supergroup)
Mesoproterozoic rocks
Nipissing gabbro
Felsic intrusions
Cobalt group rocks
Pre-Cobalt Huronian rocks
Gabbro-anorthosite
Paleoproterozoic rocks
Superior Province
Granitic and gneissic rocks
Metavolcanic rocks
Lake Superior
Canada
USA
Sault Ste. Marie
Flack Lake Fault
Quirke Syncline
Chiblow Anticline
Elliot Lake
Murray Fault System
Cutler
Bruce Mines
Thessalon
Blind River
North Channel
Lake Huron
4
5
12
N
0
40 km
84°W
83°W
Nickel-copper-PGE deposits in Nipissing diabase intrusions, southwestern Southern Province
Nickel-copper-PGE deposits in gabbro-anorthosite intrusions
Contact metasomatism deposit (no skarn assemblage)
Contact metasomatism skarn deposits
Iron-copper
Zinc-lead-copper-cobalt
Tungsten-copper-molybdenum-zinc

**Figure 23.4.** Locations of Proterozoic nickel-copper-platinum group element(PGE) mineralization that occurs in, and is associated with, 2.48 billion-year-old (Krogh et al. 1984) gabbro-anorthosite and 2.22 billion-year-old (Corfu and Andrews 1986) Nipissing gabbro. Also shown are locations of contact metasomatic deposits associated with Nipissing gabbro in Proterozoic rocks, between Sudbury and Sault Ste. Marie (*modified from* Pearson 1980). Nickel-copper-PGE deposits associated with Nipissing gabbro— 1 to 3; contact metasomatic deposits—4 to 12; and nickel-copper-PGE mineralization associated with gabbro-anorthosite intrusions—13 and 14.

Fault (*see* Figure 23.4). Northeast of the Sudbury Igneous Complex, a smaller cluster of nickel-copper-PGE occurrences is spatially associated with intrusions of Nipissing gabbro that cut Gowganda Formation rocks (*see* Figure 23.4).

Typical copper-nickel-PGE mineralization occurs within gabbroic phases of the Nipissing gabbro, although some mineralization is associated with granophyric phases and jointed areas of the intrusions (Card 1978; B. Gates, Ministry of Northern Development and Mines, Sudbury, personal communication, 1990). Mineralization exists as disseminations, lenses and pods and consists of pyrrhotite, chalcopyrite and pyrite (Pearson 1980; Innes and Colvine 1984). Most deposits contain less than 10 000 t of copper mineralization (Card and Pattison 1973; Pearson 1980), although at the Falconbridge Limited's Shakespeare Township deposit (*see* Figure 23.4), a resource of approximately 2 to 3 million t grading 0.34% Ni and 0.40% Cu has been estimated (Shklanka 1969, p.264).

At some of the deposits, sulpharsenide minerals, such as arsenopyrite and gersdorffite, are present and elevated concentrations of cobalt, platinum and palladium occur (Card 1976; Pearson 1980; Conrod 1988; Macdonald and Cherry 1988). Albitization of country rock, including the spatially associated Nipissing gabbro, is a characteristic of these copper-cobalt-sulpharsenide occurrences (cf., Gimby prospect, Pearson 1980, p.191). A genetic relationship between the secondary, hydrothermal albite and the sulpharsenide mineralization has been suggested (cf., Meyer et al. 1986, 1987, 1988, 1989, 1990; Gates 1991). For example, at



1 Falconbrige Limited's Shakespeare Township deposit
2 Rathbun Lake
3 Curtin Township
4 Two Horse Lake
5 Cobden River
6 Foster Township
7 Hess Township
8 Leger (Manitou Lake)
9 Hart Township
10 Skunk Lake
11 Sudbury ski area
12 Desbarats Lake
13 East Bull Lake
14 Shakespeare-Dunlop

**Figure 23.4.** Locations of Proterozoic nickel-copper-platinum group element(PGE) mineralization that occurs in, and is associated with, 2.48 billion-year-old (Krogh et al. 1984) gabbro-anorthosite and 2.22 billion-year-old (Corfu and Andrews 1986) Nipissing gabbro. Also shown are locations of contact metasomatic deposits associated with Nipissing gabbro in Proterozoic rocks, between Sudbury and Sault Ste. Marie (*modified from* Pearson 1980). Nickel-copper-PGE deposits associated with Nipissing gabbro— 1 to 3; contact metasomatic deposits—4 to 12; and nickel-copper-PGE mineralization associated with gabbro-anorthosite intrusions—13 and 14.

Figure 23.5. Anorthosite nodules within the mineralized inclusion-bearing leucogabbronorite unit of the East Bull Lake Intrusion. Palladium- and platinum-bearing minerals and chalcopyrite, pyrrhotite and pentlandite occur disseminated in this marginal unit of the intrusion (photograph *courtesy of* Dave Peck, Department of Geology, Laurentian University).

the Rathbun Lake prospect (*see* Figure 23.4; Pearson 1980; Dressler 1982; Finn et al. 1982), anomalous concentrations of copper, nickel, silver, gold, platinum and palladium occur in small (less than 12 m long) sulphide pods. The pods are contained entirely within, but at the lower contact of, the Wanapitei gabbronorite intrusion, where the intrusion lies in contact with siltstone of the Gowganda Formation (Macdonald and Cherry 1988). Chalcopyrite and pyrite are the abundant sulphide minerals (Macdonald and Cherry 1988). The most abundant platinum group minerals are merenskyite and kotulskite (Rowell and Edgar 1986). Some platinum occurs in sperrylite and some palladium occurs in bismuth- and telluride-bearing minerals. The sperrylite and bismuth -and telluride-bearing minerals may have formed at lower temperatures from hydrothermal fluids (Rowell and Edgar 1986). The silicate minerals of the gabbronorite are variably altered to chlorite and sericite (Dressler 1982). The origin of the mineralization has been ascribed to both magmatic (Dressler 1982) and hydrothermal (Rowell and Edgar 1986) processes.

## Nickel-Copper-Platinum Group Element Deposits Associated with the Sudbury Igneous Complex

Important deposits of magmatic nickel-copper-PGE sulphide mineralization occur along the South Range, North Range, East Range and within offset dikes of the 1.85 billion-year-old (Krogh et al. 1984) Sudbury Igneous Complex (Figure 23.6; Naldrett 1984a). This mineralization is spatially associated with the inclusion-bearing, quartz diorite Sublayer (cf., Hawley and Stanton 1962; Grant and Bite 1984). A more comprehensive discussion of these deposits is given by Dressler et al., (this volume).

Mineralization along the North Range occurs within the Sublayer, footwall breccia and adjacent footwall rocks (Coats and Snajdr 1984). There is generally a sharp contact between the mineralized Sublayer and the overlying Main Mass of quartz-rich or felsic norite (Coats and Snajdr 1984; Naldrett 1984b). Mineralization consists of pyrrhotite, chalcopyrite and pentlandite, associated with minor amounts of pyrite, magnetite and ilmenite (Coats and Snajdr 1984; Naldrett 1984b).

Locally, veins and disseminated sulphide mineralization occur in the brecciated footwall along the North Range (Coats and Snajdr 1984). This style of mineralization consists of chalcopyrite, cubanite, pyrrhotite and pentlandite, accompanied by minor amounts of sphalerite, nickel-rich mackinawite, bornite, millerite and magnetite (Coats and Snajdr 1984). This mineral assemblage is richer in, copper, platinum, palladium and gold with respect to the massive and disseminated sulphide mineralization within the Sublayer (Naldrett 1984a, 1984b).

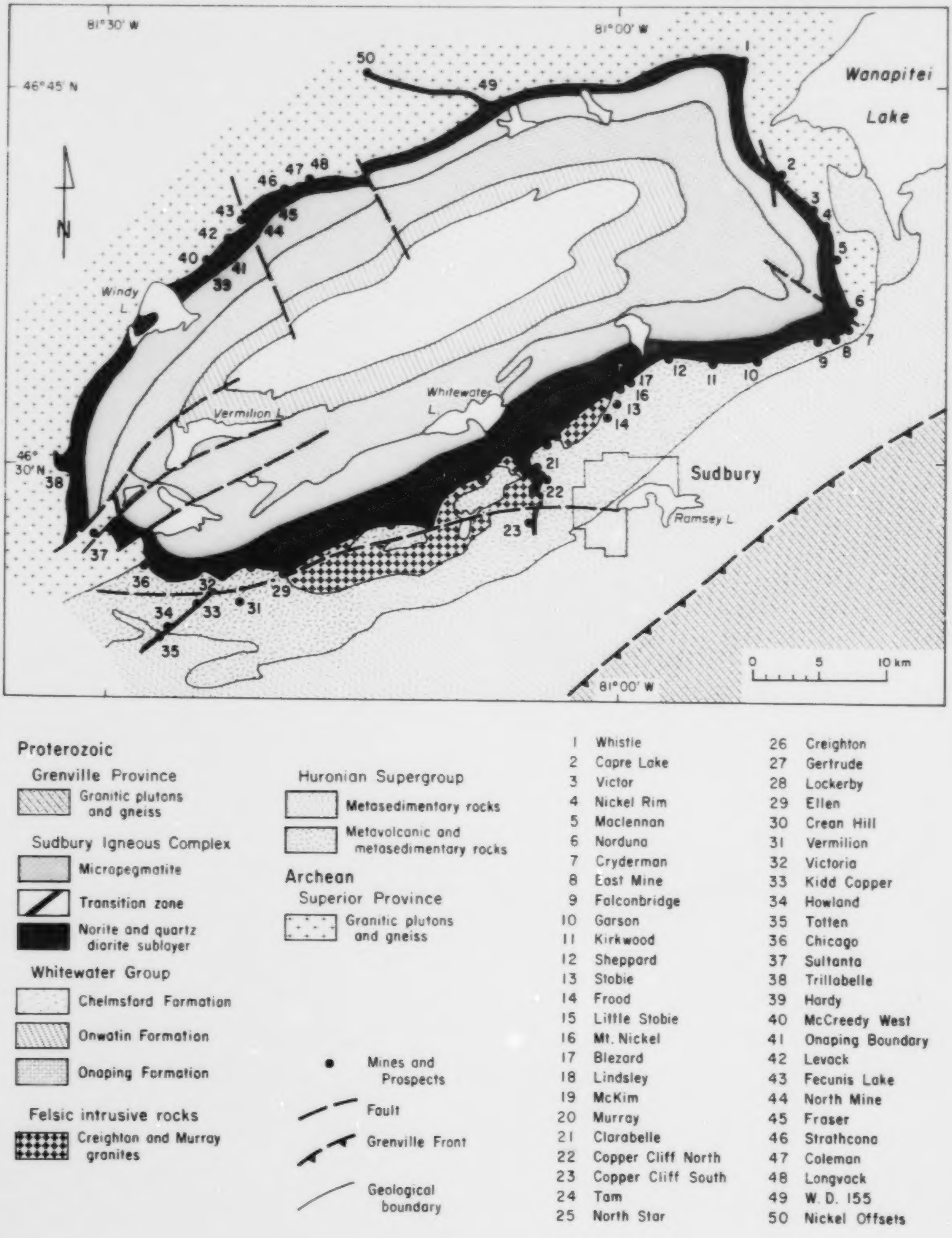
The setting of the South Range Deposits is generally similar to those of the North Range, except that the footwall is not as brecciated and the Sublayer is commonly xenolith-bearing (Naldrett 1984a). Massive, xenolith-bearing sulphide zones occur in sharp contact with the footwall rocks and the overlying quartz diorite. The quartz diorite can contain up to 60% disseminated and blebby sulphide minerals (Naldrett 1984a) and there is usually a sharp contact with the overlying Main Mass of norite (Naldrett 1984a).

The quartz diorite Sublayer forms dikes that are radially, concentrically and discontinuously distributed about the Sudbury Igneous Complex (Grant and Bite 1984). Sublayer dikes are host to many of the nickel-copper-PGE sulphide deposits in the Sudbury area (cf., Frood, Stobie, Clarabelle, Copper Cliff North, Copper Cliff South). The emplacement of the quartz diorite Sublayer was contemporaneous with, or postdated that of, the lower part of the Main Mass of norite (Grant and Bite 1984). Some phases of the dikes contain both locally derived and exotically derived gabbro and ultramafic xenoliths. Sulphide minerals occur ubiquitously as disseminated blebs, and locally as near vertical, pipe-shaped massive zones consisting of chalcopyrite, pyrrhotite and pentlandite (Grant and Bite 1984). Exotic fragments are common in the more strongly mineralized zones (Grant and Bite 1984). Sulphide liquid and xenolith fragments were incorporated into the quartz diorite magma, and transported mechanically to footwall deposit sites (Grant and Bite 1984).

Several deposits, such as the East, Falconbridge and Garson mines, are associated with faults (Naldrett 1984a). Few comprehensive descriptions exist of these fault-associated deposits.

## Nickel-Copper-Platinum Group Element Mineralization Associated with Keweenawan Tholeiitic Intrusions

The Crystal Lake Gabbro (Figure 23.7), in the Thunder Bay area, is associated with large tonnages of low-grade, nickel-copper-PGE mineralization. This layered intrusion, considered to be comparable in age with the 1.1 billion-year-old Duluth Gabbro (*see* Figure 23.2), intruded into shales of the Rove Formation (Macdonald and Cherry 1988, p.143;


81°30' W
81°00' W
46°45' N
46° 30' N
N
Wanapitei Lake
Windy L.
Vermilion L.
Whitewater L.
Sudbury
Ramsey L.
0 5 10 km
81°00' W
Proterozoic
Grenville Province
Granitic plutons and gneiss
Sudbury Igneous Complex
Micropegmatite
Transition zone
Norite and quartz diorite sublayer
Whitewater Group
Chelmsford Formation
Onwatin Formation
Onaping Formation
Felsic intrusive rocks
Creighton and Murray granites
Huronian Supergroup
Metasedimentary rocks
Metavolcanic and metasedimentary rocks
Archean
Superior Province
Granitic plutons and gneiss
Mines and Prospects
Fault
Grenville Front
Geological boundary

| | | | |
|---|---|---|---|
| 1 | Whistle | 26 | Creighton |
| 2 | Capre Lake | 27 | Gertrude |
| 3 | Victor | 28 | Lockerby |
| 4 | Nickel Rim | 29 | Ellen |
| 5 | Maclennan | 30 | Crean Hill |
| 6 | Norduna | 31 | Vermilion |
| 7 | Cryderman | 32 | Victoria |
| 8 | East Mine | 33 | Kidd Copper |
| 9 | Falconbridge | 34 | Howland |
| 10 | Garson | 35 | Totten |
| 11 | Kirkwood | 36 | Chicago |
| 12 | Sheppard | 37 | Sultanta |
| 13 | Stobie | 38 | Trillabelle |
| 14 | Frood | 39 | Hardy |
| 15 | Little Stobie | 40 | McCreedy West |
| 16 | Mt. Nickel | 41 | Onaping Boundary |
| 17 | Blezard | 42 | Levack |
| 18 | Lindsley | 43 | Fecunis Lake |
| 19 | McKim | 44 | North Mine |
| 20 | Murray | 45 | Fraser |
| 21 | Clarabelle | 46 | Strathcona |
| 22 | Copper Cliff North | 47 | Coleman |
| 23 | Copper Cliff South | 48 | Longvack |
| 24 | Tam | 49 | W. D. 155 |
| 25 | North Star | 50 | Nickel Offsets |

**Figure 23.6.** Distribution of nickel-copper-platinum group element sulphide deposits associated with the Sudbury Igneous Complex.

Sutcliffe, this volume). The mineralization at the Great Lakes Nickel deposit occurs within a pegmatitic gabbro in the middle part of the intrusion (Macdonald and Cherry 1988, p.141). The mineralization consists of disseminated chalcopyrite, pyrrhotite and pentlandite and occurs at the top of a disseminated chromite zone (Macdonald and Cherry 1988, p.143). Chromite occurs as layers and disseminated grains at the base of some anorthosite units (Macdonald and Cherry 1988, p.141). Similar mineralization occurs in the Duluth Gabbro (*see* Figure 23.2). Sulphur saturation of the parental magma is attributed to assimilation of sulphur-rich sedimentary rock (cf., Ripley 1981).

### Nickel-Copper-Platinum Group Element Mineralization Associated with Keweenawan Dikes

In the Thunder Bay area, disseminated nickel-copper-PGE sulphide mineralization occurs within biotite-bearing, olivine gabbro Keweenawan dikes that extend from Pigeon Point to Edward Island (*see* Figure 23.7; Franklin 1970). At Pigeon Point (*see* Figure 23.7), the mineralization consists of disseminated pyrrhotite, chalcopyrite and pentlandite associated with ilmenite (Franklin 1970). The mineralization occurs generally within, but along the margin of the dikes (Franklin 1970). At the Edward Island occurrence (*see* Figure 23.7), chalcopyrite, pentlandite, pyrrhotite and pyrite occur as ovoid sulphide segregations (Franklin 1970). Within the host gabbro dike, feldspar is albitized, and pyroxene is altered to chlorite (Franklin 1970).

### Nickel-Copper-PGE Mineralization Associated with Keweenawan Alkalic Intrusions

Three copper-nickel-PGE occurrences are localized within the Port Coldwell alkalic complex (*see* Figure 23.7). A copper and platinum group element (PGE) deposit occurs in coarse-grained to pegmatitic gabbro, on the eastern margin of the Port Coldwell alkalic complex, in the Bamoos Lake area (cf., Macdonald and Cherry 1988; *see* Sutcliffe, this volume). The mineralization consists of disseminated chalcopyrite, cubanite, pyrrhotite, pentlandite and pyrite, has a high Pd-Pt ratio (~10/1) and contains traces of nickel, cobalt, rhodium and silver (Dahl et al. 1986). Sulphur-saturation of the magma is attributed to the assimilation of sulphur-rich Archean rock (Dahl et al. 1986). Other copper-gold-nickel-PGE occurrences include the Coubran Lake, Middleton and MacRae. The MacRae occurrence is found within the interior of the Port Coldwell alkalic complex, along a gabbro-syenite contact (*see* Sutcliffe, this volume).

## URANIUM MINERALIZATION

### Paleoplacer Uranium—Huronian Supergroup

The uranium deposits in the Elliot Lake area are considered to be paleoplacer in origin (Roscoe 1969; Robertson 1986, *see* Bennett et al., this volume). Pyrite, uraninite and brannerite occur within the lower Matinenda Formation, Elliot Lake Group, at the base of the Huronian Supergroup (cf., Robertson 1978). The pyrite and uraninite are considered to be detrital in origin. Associated brannerite is interpreted to have formed during diagenetic remobilization and adsorption of dissolved uranium on titaniferous minerals (cf., Roscoe 1969; Theis 1978). Mineralization displays a strong control by sedimentary structures and facies, showing a close relationship between mineralization tenor, U-Th ratios, pebble size and heavy mineral concentrations, consistent with a placer origin (cf., Roscoe 1969; Theis 1978).

### Unconformity-Related Uranium

Uranium mineralization occurs in several settings in the Thunder Bay area (Carter 1977; Fenwick and Scott 1977; Robertson 1978): 1) uraninite in Archean pegmatites that cut migmatitic metasedimentary rocks of the Quetico Subprovince; 2) pyrite-pitchblende-quartz veins in joints in quartz-monzonite intrusions; 3) uraniferous marcasite-pyrite veins in the northwest-striking Christianson fault; 4) coffinite near the unconformity between the Sibley Group and Archean rocks (*see* Figure 23.7); and 5) uraniferous minerals in the black shale of the Rove Formation and in tuffaceous shale and arenite of the Gunflint Formation near Kakabeka Falls.

The uranium mineralization that is spatially associated with quartz-monzonite and the Christianson fault is post-Archean in age. Although the uranium veins occur in faults that strike 330°, the veins are most frequently developed along and near the unconformity between the Sibley Group and the Archean basement (*see* Figure 23.7; Franklin 1970; Franklin et al. 1982). The fault localized mineralization formed during the Proterozoic and may be related to the development of the Sibley Basin (Yule 1979).

## ANOROGENIC GRANITE MINERALIZATION

The presence of the 1.5 billion-year-old (Davis and Sutcliffe 1985) anorogenic (A-type) English Bay Granite in the Lake Nipigon area suggests that disseminated, vein, pegmatite or greisen-related beryllium, niobium, tantalum, yttrium, REE, fluorine, zirconium, thorium, tin, tungsten and molybdenum mineralization may also occur (cf., Martin and Bowden 1981). Although fluorite and biotite are present in the English Bay Granite, no metallic mineralization has yet been discovered.

## SEDIMENT-HOSTED, RED-BED, STRATABOUND COPPER MINERALIZATION

Red-bed, stratabound copper mineralization occurs within red-coloured sedimentary successions in the upper part of the Huronian Supergroup, in the Elliot Lake area, at the base of the Sibley Group and in Upper Keweenawan sedimentary rocks, in the Lake Superior area. This type of mineralization

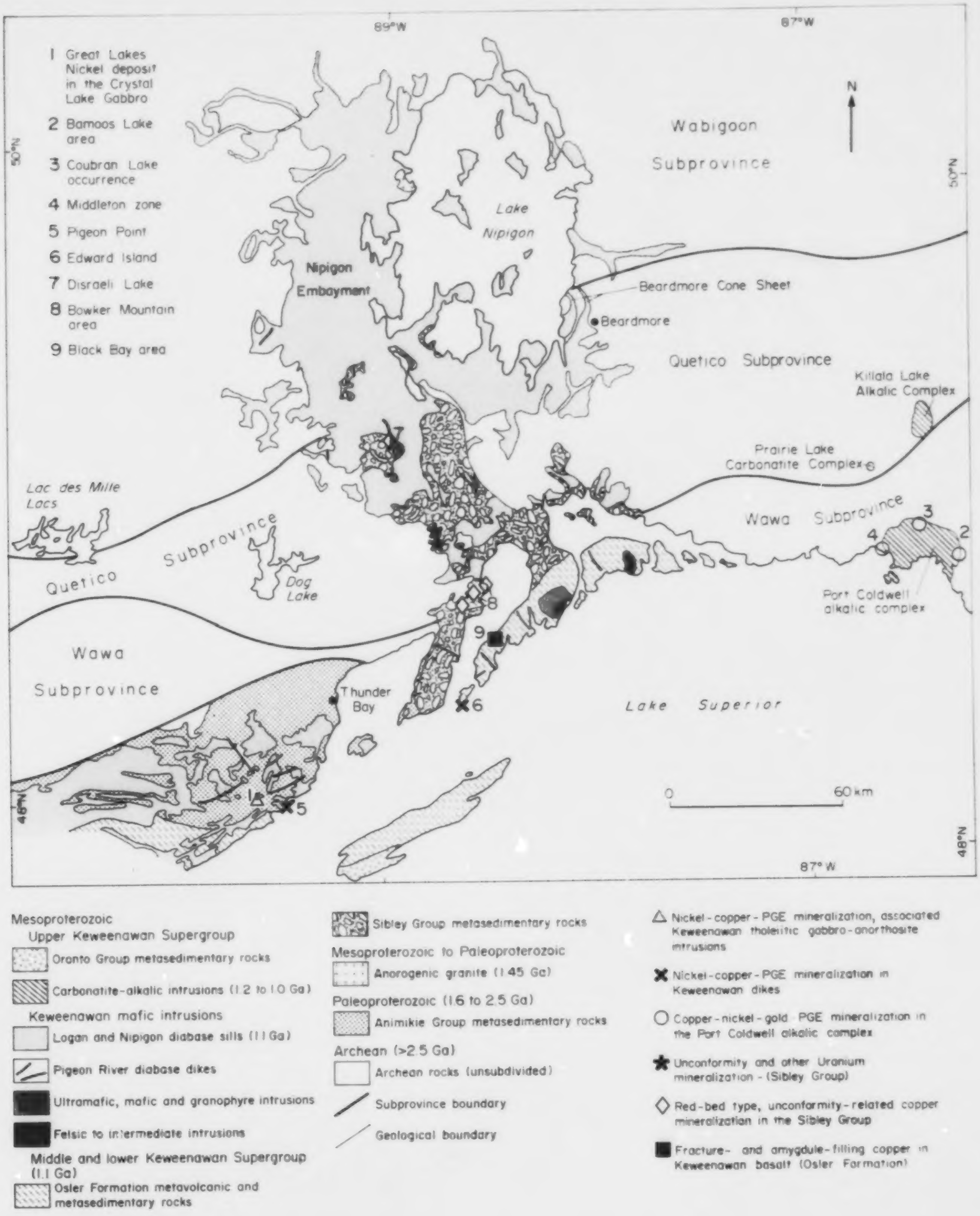

89°W
87°W
50°N
48°N
87° W
N
1 Great Lakes Nickel deposit in the Crystal Lake Gabbro
2 Bamoos Lake area
3 Coubran Lake occurrence
4 Middleton zone
5 Pigeon Point
6 Edward Island
7 Disraeli Lake
8 Bowker Mountain area
9 Black Bay area
Wabigoon Subprovince
Lake Nipigon
Nipigon Embayment
Beardmore Cone Sheet
Beardmore
Quetico Subprovince
Killala Lake Alkalic Complex
Prairie Lake Carbonatite Complex
Lac des Mille Lacs
Wawa Subprovince
Quetico Subprovince
Dog Lake
Wawa Subprovince
Thunder Bay
Port Coldwell alkalic complex
Lake Superior
0
60 km
Mesoproterozoic
Upper Keweenawan Supergroup
Oronto Group metasedimentary rocks
Carbonatite-alkalic intrusions (1.2 to 1.0 Ga)
Keweenawan mafic intrusions
Logan and Nipigon diabase sills (1.1 Ga)
Pigeon River diabase dikes
Ultramafic, mafic and granophyre intrusions
Felsic to intermediate intrusions
Middle and lower Keweenawan Supergroup (1.1 Ga)
Osler Formation metavolcanic and metasedimentary rocks
Sibley Group metasedimentary rocks
Mesoproterozoic to Paleoproterozoic
Anorogenic granite (1.45 Ga)
Paleoproterozoic (1.6 to 2.5 Ga)
Animikie Group metasedimentary rocks
Archean (>2.5 Ga)
Archean rocks (unsubdivided)
Subprovince boundary
Geological boundary
△ Nickel-copper-PGE mineralization, associated Keweenawan tholeiitic gabbro-anorthosite intrusions
✖ Nickel-copper-PGE mineralization in Keweenawan dikes
○ Copper-nickel-gold PGE mineralization in the Port Coldwell alkalic complex
★ Unconformity and other Uranium mineralization - (Sibley Group)
◇ Red-bed type, unconformity-related copper mineralization in the Sibley Group
■ Fracture- and amygdule-filling copper in Keweenawan basalt (Osler Formation)

**Figure 23.7.** Distribution of metallic mineral deposit types in the Lake Superior area of Ontario. Nickel-copper-platinum group element (PGE) mineralization associated with Keweenawan tholeiitic gabbro-anorthosite intrusions— 1; Nickel-copper-platinum group element mineralization associated with the Port Coldwell alkalic complex—2 to 4; Nickel-copper-platinum group element mineralization associated with Keweenawan dikes—5 and 6; red-bed copper mineralization in the Sibley Group—7 and 8; and fracture- and amygdule-filling copper in Osler Formation basalts—9.

consists of disseminated pyrite, chalcopyrite, bornite and chalcocite localized within, and adjacent to, reduced shale units, in proximity to red-coloured arenite units (Kirkham 1989). On a worldwide basis, virtually all red-bed copper deposits occur in rocks younger than 1.3 Ga (Kirkham 1989; Chandler 1989). The deposits and occurrences in the Huronian Supergroup are, therefore, unusual because they occur in sedimentary rocks that are approximately 2.4 billion years old.

### Red-bed Copper Mineralization in the Huronian Supergroup

Within the Huronian Supergroup, red-bed copper mineralization occurs in alluvial, white-weathering, quartz arenites at the top of the Lorrain Formation, Cobalt Group (*see* Figure 23.3; Pearson 1980; Chandler 1989). These arenites are overlain by a marine arenite-mudstone unit of the Gordon Lake Formation (Chandler 1989). Along this contact, sabkha-related clastic metasedimentary rocks, with zones of anhydrite nodules, also occur (Chandler 1989). Thin zones of red beds occur above and below the sabkha sedimentary rocks (Chandler 1989).

Although disseminated pyrite and chalcopyrite occur throughout the Lorrain Formation, the greatest concentrations of chalcopyrite occur in the sabkha-related red beds at the Stag Lake and Cobre Lake occurrences (*see* Figure 23.3; Frarey 1977; Pearson 1980; Chandler 1989). Here, the mineralization consists of disseminated, veinlet and nodular chalcopyrite, pyrite and rare bornite (Pearson 1980; Chandler 1989). At the Stag Lake occurrence, copper tenor is about 0.2% Cu, and the mineralized zone is generally much less than 10 m in thickness (Pearson 1980). At the Stag Lake occurrence (*see* Figure 23.3), the host mineralized rock of the Lorrain Formation is slightly radioactive and contains hematitic, calcareous, pink feldspathic and sericitic quartzite, in proximity to overlying green mudstone and siltstone units of the Gordon Lake Formation (Pearson 1980). Some hematite at Cobre Lake is interpreted to have formed diagenetically (Chandler 1989). Generally, the chalcopyrite-rich areas are associated with disseminated specular hematite, and the host rocks are extensively albitized (Pearson 1980). Textural evidence is consistent with a diagenetic origin of the disseminated pyrite (Chandler 1989).

Pyrite-chalcopyrite mineralization also occurs at a few locations within the basal arkose member of the Lorrain Formation, Cobalt Group, such as at the Desbarats occurrence. In the Cobalt area, conglomerates and siltstone of the Gowganda Formation are also host to disseminated base metal mineralization (Chandler 1989), although no comprehensive descriptions of this mineralization exist (Innes and Colvine 1984). These stratabound concentrations of pyrite and chalcopyrite are also interpreted to be red-bed copper deposits (Chandler 1989).

Formation of pyrite and chalcopyrite, in the Huronian sabkha sedimentary rocks, is attributed to the reduction and precipitation of chalcopyrite from oxidized, copper-bearing solutions that flowed through permeable arkose and quartz arenite units. The hydrothermal fluids reacted with reducing green siltstone or mudstone units of the overlying Gordon Lake Formation (Rose 1976; Pearson 1980). Alternatively, the presence of gypsum, at some localities, as nodules, veins and cement at the base of the Gordon Lake Formation (cf., Wood 1973; Chandler 1986) indicates a sabkha depositional origin (Chandler 1989).

### Red-bed Copper Mineralization in the Lake Superior Area

Historically, important red-bed, stratabound copper mineralization, typified by that at the White Pine Mine, occurs in the postvolcanic, Upper Keweenawan sediments of the Oronto Group in Michigan (Ensign et al. 1968). This mineralization occurs in 4 laterally extensive shale and siltstone units that overly copper-poor sandstone within the basal part of the Nonesuch Shale (cf., Ensign et al. 1968; Brown 1971; Seasor and Brown 1989). The mineralization consists principally of disseminated chalcocite, digenite, bornite, chalcopyrite and lesser amounts of native copper (Ensign et al. 1968; Brown 1971). Native copper veins occur in the upper part of the fractured Copper Harbour Conglomerate, stratigraphically below the Nonesuch Shale (Ensign et al. 1968). The Freda Formation, stratigraphically overlying the Nonesuch Shale, is a fluvial red-bed arenite but is not mineralized. In Ontario, rocks of the Jacobsville Formation are interpreted to unconformably overlie the Oronto Group and occur at the Mamainse Point area, eastern Lake Superior (*see* Sutcliffe, this volume). Folded sandstone units near Mamainse Point, Ontario, have been correlated with the Freda Formation (cf., Giblin 1969), the upper part of the Oronto Group. In this area, chalcocite-rich mineralization occurs in quartz-calcite veins, accompanied by lesser amounts of bornite, chalcopyrite and native copper which cut Keweenawan basalt of the Mamainse Point Formation (Giblin 1966). Sediment-hosted, stratabound copper mineralization of the White Pine type has not yet been discovered in this area in Ontario.

Copper mineralization of possible red-bed type occurs locally within and along the basal unconformity of the Sibley Group (*see* Figure 23.7). Mineralization consists of disseminated chalcocite, chalcopyrite, galena, sphalerite, marcasite and pyrite (Franklin 1970). At Disraeli Lake, disseminated and vug-filling digenite, cuprite, covellite, chalcopyrite, native copper and malachite occur in a stromatolitic unit, in close proximity to a mafic intrusion (Franklin 1970). The Bowker area copper showings consist of chalcopyrite-chalcocite-quartz-pyrite veins in the Sibley red mudstone at, or near, the unconformity with the Archean basement (Franklin 1970).

## FRACTURE- AND AMYGDULE-FILLING COPPER IN BASALT

In the Lake Superior area, chalcocite, covellite, quartz, calcite and native copper fill amygdules and fractures in columnar-jointed basalt flows near the base of the Osler Formation (*see* Figure 23.7; Franklin 1970). Vesicles are also filled by zeolite minerals, epidote and chlorite (Franklin

1970). Native copper and chalcocite also occur as disseminations, along bedding planes, and in fractures, within red-coloured wacke units, interlayered within the volcanic rocks of the Osler Formation (Franklin 1970). This style of copper mineralization is similar to the copper deposits in the Keweenaw Peninsula of Michigan (Franklin 1970). Here, native copper, chalcocite, covellite and bornite, associated with zeolite minerals, chlorite, epidote and calcite, occur primarily in conglomerate beds, amygdaloidal flow tops and less commonly as veins within the Portage Lake Volcanics (Lindgren 1933; Weege and Pollock 1971; Franklin et al. 1982). The Portage Lake Volcanics underlies the copper-bearing Nonesuch Shale.

## CONTACT METASOMATIC MINERALIZATION

In the Huronian Supergroup, contact metasomatic deposits occur along the contacts of Nipissing gabbro (*see* Figure 23.4). Two types of deposits are recognized (Pearson 1980): 1) skarn assemblages in the calcareous Espanola Formation; and 2) contact metasomatic deposits in noncalcareous sedimentary rocks.

Contact metasomatic deposits are few in number and appear to be small in size (Pearson 1980). At the Desbarats Lake occurrence (*see* Figure 23.4), disseminated chalcopyrite and pyrite occur in quartz veins within quartzite of the Lorrain Formation along the contact of a Nipissing gabbro (Pearson 1980). Radioactivity associated with this occurrence exceeds background by up to 500 times (Pearson 1980). Other occurrences of this type contain small lenses and disseminations of pyrrhotite and chalcopyrite (Pearson 1980) and are found within feldspathic quartzite and siltstone of the McKim Formation, near Sudbury (Pearson 1980).

About 10 sulphide-bearing skarn occurrences are located where Nipissing gabbro has intruded the carbonate-bearing rocks of the Espanola Formation (*see* Figure 23.4). Most occurrences are magnetite rich with subordinate chalcopyrite. Based on metal assemblages, 3 subtypes are recognized (Pearson 1980): 1) iron and copper; 2) zinc, lead, copper and cobalt; and 3) tungsten, copper, molybdenum and zinc. Most mineralized zones occur as massive pods, veins and disseminations within several metres of the contacts with Nipissing gabbro.

### Iron-Copper Skarns

The iron-copper skarns are magnetite rich with lesser chalcopyrite and pyrite. Chalcopyrite occurs preferentially with the poorly layered parts of the skarn, where iron-rich chlorite is abundant. The magnetite-rich parts are finely layered and contain only minor chlorite and chalcopyrite (Pearson 1980). Sphalerite, pyrite, smaltite and cobaltite can occur as accessory minerals (Pearson 1980).

### Zinc-Lead-Copper-Cobalt Skarns

The zinc-lead-copper-cobalt skarns are less abundant and contain veins, disseminations and poddy concentrations of magnetite, pyrite, sphalerite, galena, chalcopyrite and cobaltite or smaltite (Pearson 1980). Deposits can be zoned from magnetite rich, adjacent to the contact with Nipissing gabbro, outwards through to chalcopyrite-pyrite-, sphalerite-galena- and cobalt-bearing assemblages (Pearson 1980).

### Tungsten-Copper-Molybdenum-Zinc Skarns

Few tungsten-copper-molybdenum-zinc skarns are known. The Foster Township occurrence is a diopside-garnet-idocrase skarn, which lacks magnetite (Pearson 1980). It is characterized by up to 1% fracture-controlled and disseminated scheelite, associated with pyrrhotite, pyrite, chalcopyrite, molybdenite and sphalerite (Card 1976; Pearson 1980).

## HYDROTHERMAL VEINS

Several types of hydrothermal vein assemblages occur in Proterozoic rocks of the northern Great Lakes area (*see* Figures 23.8 and 23.9). A six-fold classification, based on metal assemblages, has been proposed (Franklin 1970; Innes and Colvine 1979; Pearson 1980): 1) quartz-carbonate-chalcopyrite-pyrite-pyrrhotite-specularite (copper); 2) sphalerite-galena-chalcopyrite-pyrite (zinc-lead-silver-copper); 3) chalcopyrite-cobaltite (copper-cobalt-[bismuth]-[uranium]-[nickel]); 4) bornite-chalcocite-native copper; 5) gold-quartz-arsenopyrite-pyrite-chalcopyrite; and 6) sulpharsenide (silver-cobalt-arsenic). Of these, the most economically important are the silver-cobalt sulpharsenide veins (*see* Figure 23.9), which occur in the Cobalt, Gowganda, Elk Lake and South Lorrain camps in the Cobalt Embayment (cf., Andrews et al. 1986) and in the Thunder Bay area (Franklin et al. 1986; Kissin and Jennings 1987; Kissin and Sherlock 1989). In the Huronian Supergroup, vein types "1", "2", "3", "5" and "6" strike parallel to structures that formed during the 1.86 to 1.83 billion-year-old Penokean Orogeny. In the Lake Superior area, vein types "6" and "4" formed within 1.1 billion-year-old Keweenawan faults.

In the Huronian Supergroup, regional metal distribution patterns are developed (Card and Pattison 1973; Innes and Colvine 1979, 1984). Copper-bearing veins (chalcopyrite and bornite) are more abundant in the western area, along the north shore of Lake Huron and occur between Elliot Lake and Sault Ste. Marie (Pearson 1980). A few of these are known to contain over 1000 t of copper ore and only 4 contain between 10 000 and 100 000 t of copper ore (Pearson 1980). Among these are the Bruce Mines deposits, where mining operations began in 1846 (Figure 23.10). These were the first copper deposits to be successfully mined in Canada. Polymetallic, base- and precious-metal veins increase in frequency towards the northeast (Innes and Colvine 1979, 1984). Silver-cobalt sulpharsenide veins predominate in the extreme northeast. Metal distribution patterns are also developed in the silver-bearing veins near Thunder Bay (Franklin 1970). The Mainland area veins contain acanthite, pyrite, sphalerite, marcasite, galena, calcite, quartz, barite and fluorite (Franklin et al. 1986). To the southeast are the Island veins which consist of silver-cobalt-nickel-arsenic-bismuth sulpharsenide minerals and are similar to those of the Cobalt camp (Oja 1967). A third vein array, consisting of lead-zinc-barite mineral assemblages, occurs northeast of Thunder Bay (Franklin 1970).

Within the Huronian Supergroup, the Gowganda Formation and Nipissing gabbro are the preferred vein host rocks. However, deposit size is not dependent on proximity to Nipissing gabbro, nor are areas underlain by abundant Nipissing gabbro necessarily the locus of many veins (Pearson 1980). Other vein sets, such as those that occur in the highly deformed Sudbury–Cutler segment of the Penokean Fold Belt, are spatially associated with Salmay Lake Formation metavolcanic rocks and not with Nipissing gabbro (Pearson 1980). The origin of these veins may be less related genetically to the Nipissing gabbro and more dependent on factors related to local strain state and thermal metamorphism (Pearson 1980). In the Thunder Bay area, most veins occur in the Rove Formation, in proximity to, or within, diabasic Logan sills or northeast-striking gabbro dikes (Franklin 1970).

In the Huronian Supergroup, veins strike either westerly or northwesterly (Pearson 1980). Vein types "1", "2", "3" and "5" strike westerly, parallel to major structures that developed during the Penokean Orogeny (Pearson 1980), consistent with vein formation during, or after, the Penokean Orogeny. In the Sault Ste. Marie area, bornite-chalcocite-native copper veins (type "4") strike northwesterly, parallel to Keweenawan structures consistent with vein formation during, or following, Keweenawan extension (Pearson 1980). In the Thunder Bay area, vein arrays generally strike northeasterly, parallel to regional Keweenawan faults, also consistent with vein formation during Keweenawan extension (Franklin 1970; Franklin et al. 1986).

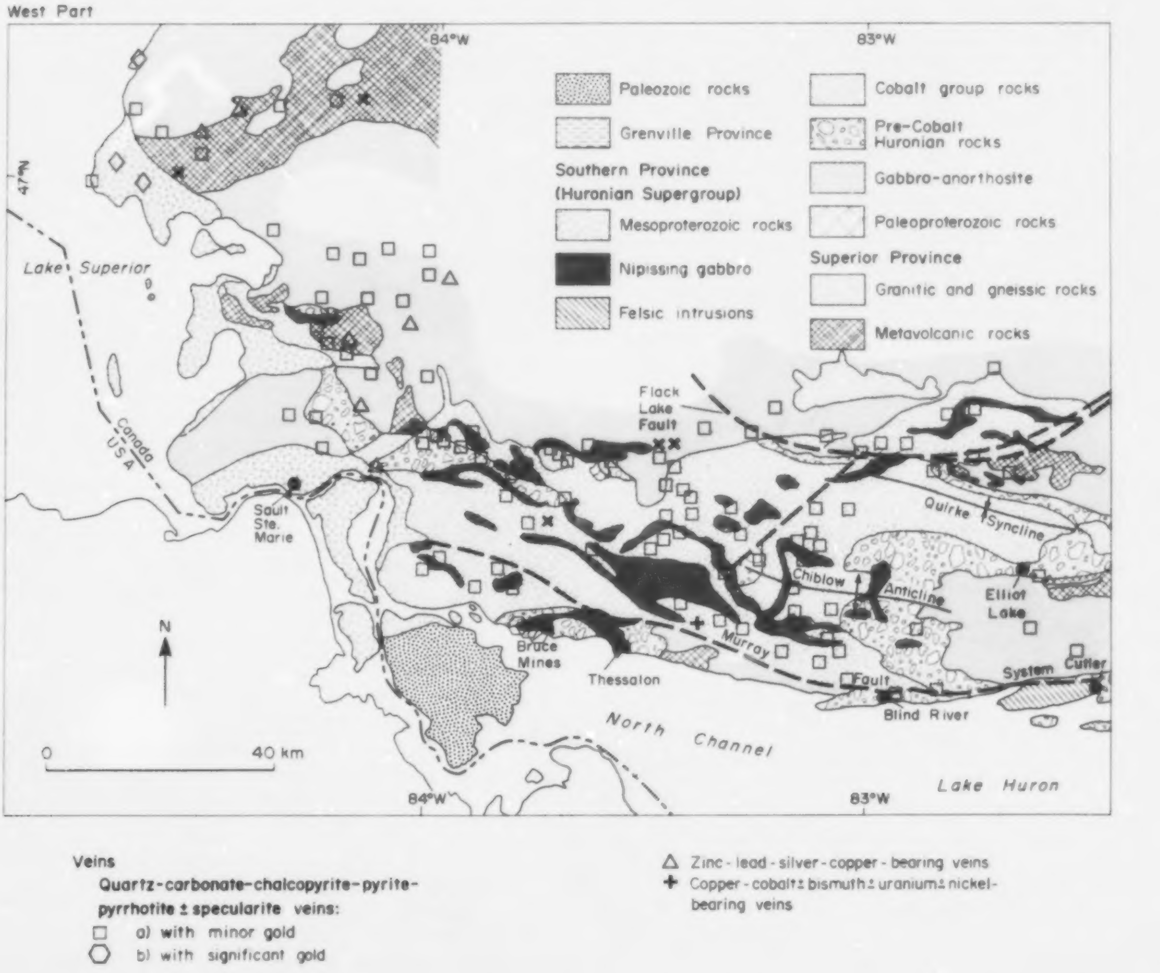

West Part
84°W
83°W
47°N
Paleozoic rocks
Grenville Province
Southern Province
(Huronian Supergroup)
Mesoproterozoic rocks
Nipissing gabbro
Felsic intrusions
Cobalt group rocks
Pre-Cobalt Huronian rocks
Gabbro-anorthosite
Paleoproterozoic rocks
Superior Province
Granitic and gneissic rocks
Metavolcanic rocks
Lake Superior
Canada
USA
Sault Ste. Marie
Flack Lake Fault
Quirke Syncline
Chiblow Anticline
Elliot Lake
Murray Fault System
Cutler
Bruce Mines
Thessalon
Blind River
North Channel
Lake Huron
N
0
40 km
Veins
Quartz-carbonate-chalcopyrite-pyrite-pyrrhotite ± specularite veins:
a) with minor gold
b) with significant gold
Zinc-lead-silver-copper-bearing veins
Copper-cobalt ± bismuth ± uranium ± nickel-bearing veins

**Figure 23.8.** Distribution of hydrothermal veins in the southwestern part of the Huronian Supergroup (*modified from* Pearson 1980). The circled area located east of the Sudbury Basin encloses numerous zones of sodium metasomatism and the associated Scadding gold and Norstar gold-copper mines. Other zones of sodium metasomatism occur elsewhere in Proterozoic rocks in the area east of Lake Huron (e.g., Meyer et al. 1987).

## Quartz, Carbonate, Chalcopyrite, Pyrite, Pyrrhotite and Specularite Veins

This is the most abundant vein assemblage in the Huronian Supergroup. These veins occur more frequently in the Elliot Lake–Blind River area (*see* Figure 23.8) and, to a lesser extent, near the western margin of the Cobalt Embayment. Host rocks are generally Gowganda Formation and Nipissing gabbro (Moore 1956; Pearson 1980). The veins dip steeply, strike easterly and consist principally of quartz, carbonate, chalcopyrite, pyrite and specularite (Figure 23.11). Specular hematite is the most abundant mineral in many deposits (Frarey 1977). The veins are commonly brecciated and are associated with secondary albite. Bornite and chalcocite are present in a few deposits and pyrrhotite is prominent in more metamorphosed veins (Pearson 1980). Gold tenor of these veins is low, except in veins located east and southeast of Wanapitei Lake (Pearson 1980; Dressler 1982; Meyer et al. 1988; Gates 1991).

## Sphalerite, Galena, Chalcopyrite and Pyrite Veins

Veins of this type within the Huronian Supergroup occur principally in the Sault Ste. Marie area (*see* Figure 23.8), where they are hosted within rocks of the Gowganda Formation (Pearson 1980). Vein mineralogy consists of galena, sphalerite, chalcopyrite, pyrite and minor pyrrhotite, specularite, magnetite, native gold and arsenopyrite (Pearson 1980). Silver can occur in solid solution in galena, as native silver and in argentite, tetrahedrite and cyprargyrite (stromeyerite-CuAgS) (Pearson 1980).

Base metal-bearing veins of this type also occur in the Thunder Bay area. Minor production was attained from galena-sphalerite-barite veins which crop out northeast of Thunder Bay, between Pass Lake and Kama Bay (Franklin 1970). These veins, associated with fault breccia zones, are localized within both Sibley Group and Archean rocks, close to the basal unconformity of the Sibley Group (Hawley 1930; Franklin 1970). Vein mineralogy consists of

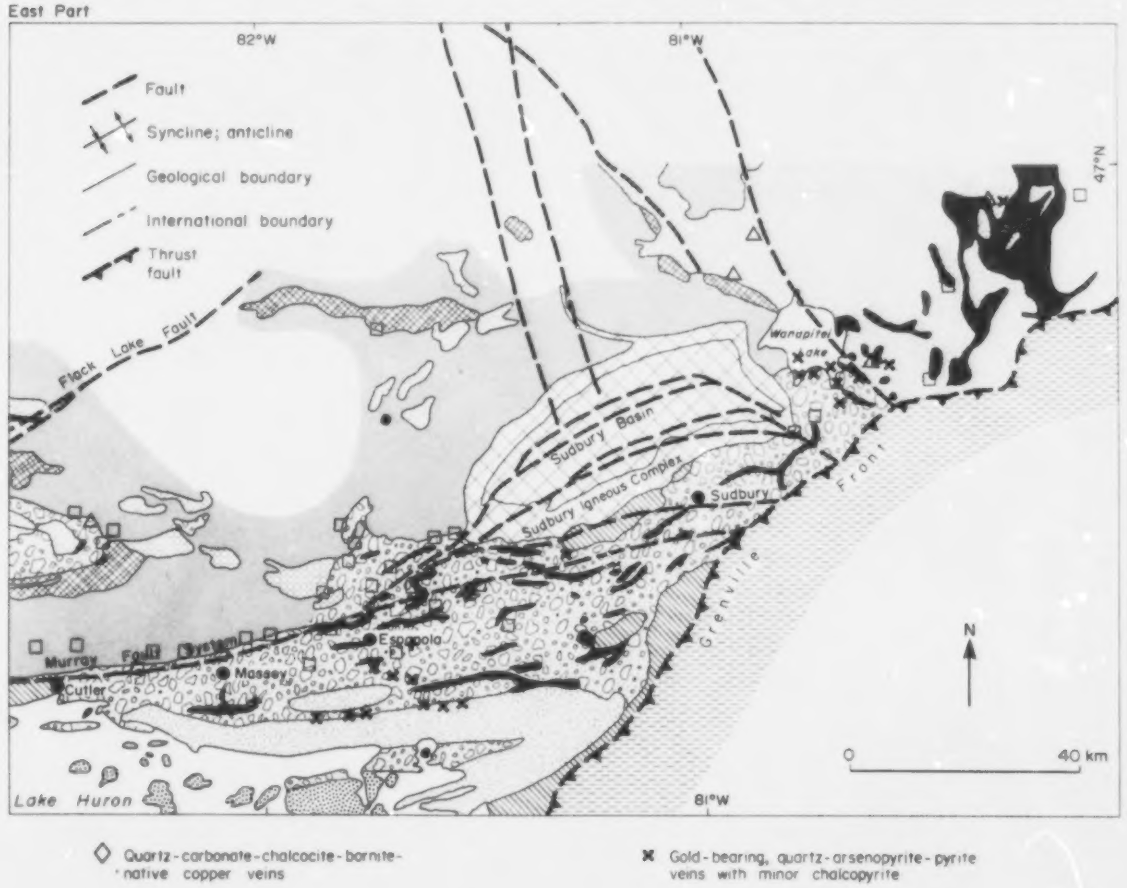

East Part
82°W
81°W
47°N
Fault
Syncline; anticline
Geological boundary
International boundary
Thrust fault
Flack Lake Fault
Wanapitei Lake
Sudbury Basin
Sudbury Igneous Complex
Sudbury
Grenville Front
Espanola
Massey
Cutler
Murray Fault System
N
0
40 km
Lake Huron
81°W
◇ Quartz-carbonate-chalcocite-bornite-native copper veins
✖ Gold-bearing, quartz-arsenopyrite-pyrite veins with minor chalcopyrite

**Figure 23.8.** Distribution of hydrothermal veins in the southwestern part of the Huronian Supergroup (*modified from* Pearson 1980). The circled area located east of the Sudbury Basin encloses numerous zones of sodium metasomatism and the associated Scadding gold and Norstar gold-copper mines. Other zones of sodium metasomatism occur elsewhere in Proterozoic rocks in the area east of Lake Huron (e.g., Meyer et al. 1987).

galena, sphalerite, pyrite, marcasite, chalcopyrite and abundant barite, quartz, calcite and minor fluorite (Hawley 1930; Franklin 1970). The fault breccia consists of granite and Sibley mudstone fragments, cemented by quartz, calcite or sulphide minerals (Franklin 1970). The Sibley mudstone, which is present as red mudstone in most occurrences in the area, is altered when adjacent to the veins, to a dolomite- and calcite-bearing silicified rock (Franklin 1970). Lead isotope data are consistent with a lead source from both Archean granitic and Sibley Group rocks (Franklin 1970).

## Chalcopyrite and Cobaltite Veins

Chalcopyrite-cobaltite veins (copper-cobalt-±bismuth± uranium±nickel) do not appear to show a specific geographic distribution in the Huronian Supergroup (*see* Figure 23.8). Like most of the vein types, they are spatially associated with Nipissing gabbro (Pearson 1980). Vein mineralogy consists of chalcopyrite, cobaltite and minor amounts of arsenopyrite, pyrite, pyrrhotite, specularite and native gold (Pearson 1980).

## Quartz, Carbonate, Chalcocite, Bornite and Native Copper Veins

Quartz-carbonate-chalcocite-bornite-native copper veins (*see* Figure 23.8) are confined to the Keweenawan volcanic and sedimentary rocks and to Archean rocks, adjacent to the Keweenawan rift zone, in the western part of the Huronian Supergroup (Giblin 1966; Pearson 1980). These veins contain chalcocite, chalcopyrite, bornite, specularite, native copper, covellite, digenite, pyrite and tetrahedrite (Pearson 1980). The parallelism between vein strike and

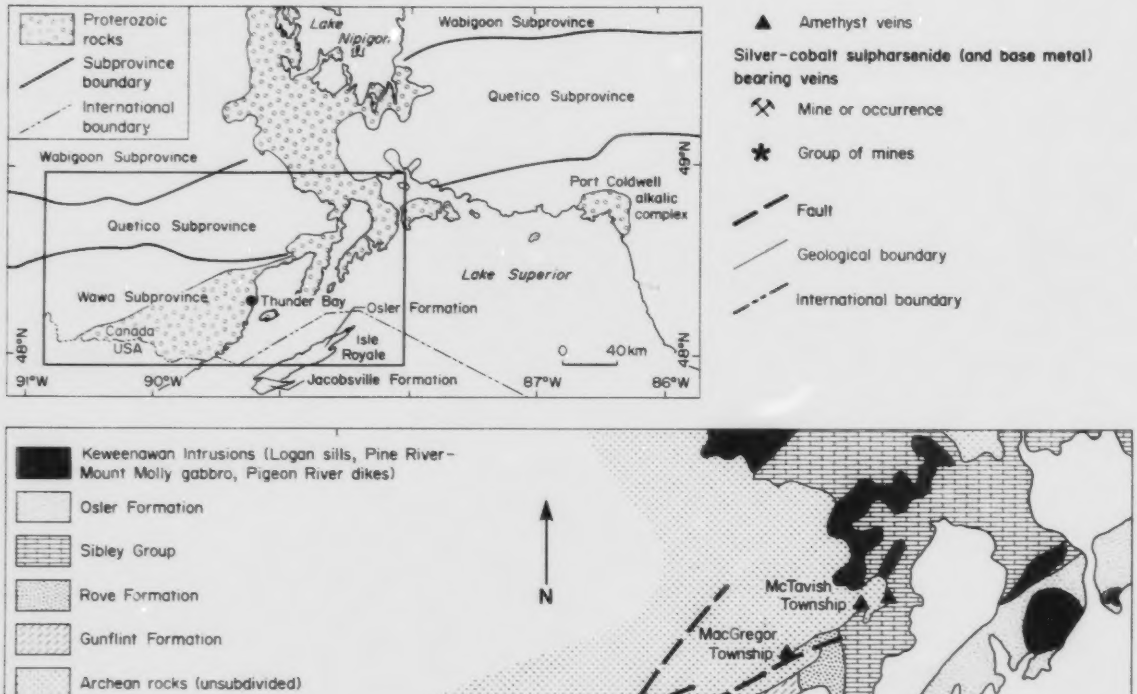

Proterozoic rocks
Subprovince boundary
International boundary
Lake Nipigon
Wabigoon Subprovince
Quetico Subprovince
Wabigoon Subprovince
Quetico Subprovince
Port Coldwell alkalic complex
Lake Superior
Wawa Subprovince
Thunder Bay
Osler Formation
Canada
USA
Isle Royale
Jacobsville Formation
0 40km
49°N
48°N
91°W
90°W
87°W
86°W
▲ Amethyst veins
Silver-cobalt sulpharsenide (and base metal) bearing veins
Mine or occurrence
Group of mines
Fault
Geological boundary
International boundary
Keweenawan Intrusions (Logan sills, Pine River-Mount Molly gabbro, Pigeon River dikes)
Osler Formation
Sibley Group
Rove Formation
Gunflint Formation
Archean rocks (unsubdivided)
N
0 50km
McTavish Township
MacGregor Township
Beck
Shuniah
Thunder Bay
Island Array
Lake
Superior
Rabbit Mountain
Array
Thunder Bay
Edward Island
Lily of the Valley
Mainland
Silver Islet
Burnt Island
Pie Island
Spar Island
Victoria Island
Canada
USA
Ontario
Minnesota
Isle Royale
89°W
48°N

**Figure 23.9. a)** Distribution of silver-cobalt sulpharsenide and amethyst veins in the Thunder Bay area and **b)** silver-cobalt sulpharsenide veins within the Cobalt Embayment. *Modified from* Andrews et al. (1986) and Franklin et al. (1986).

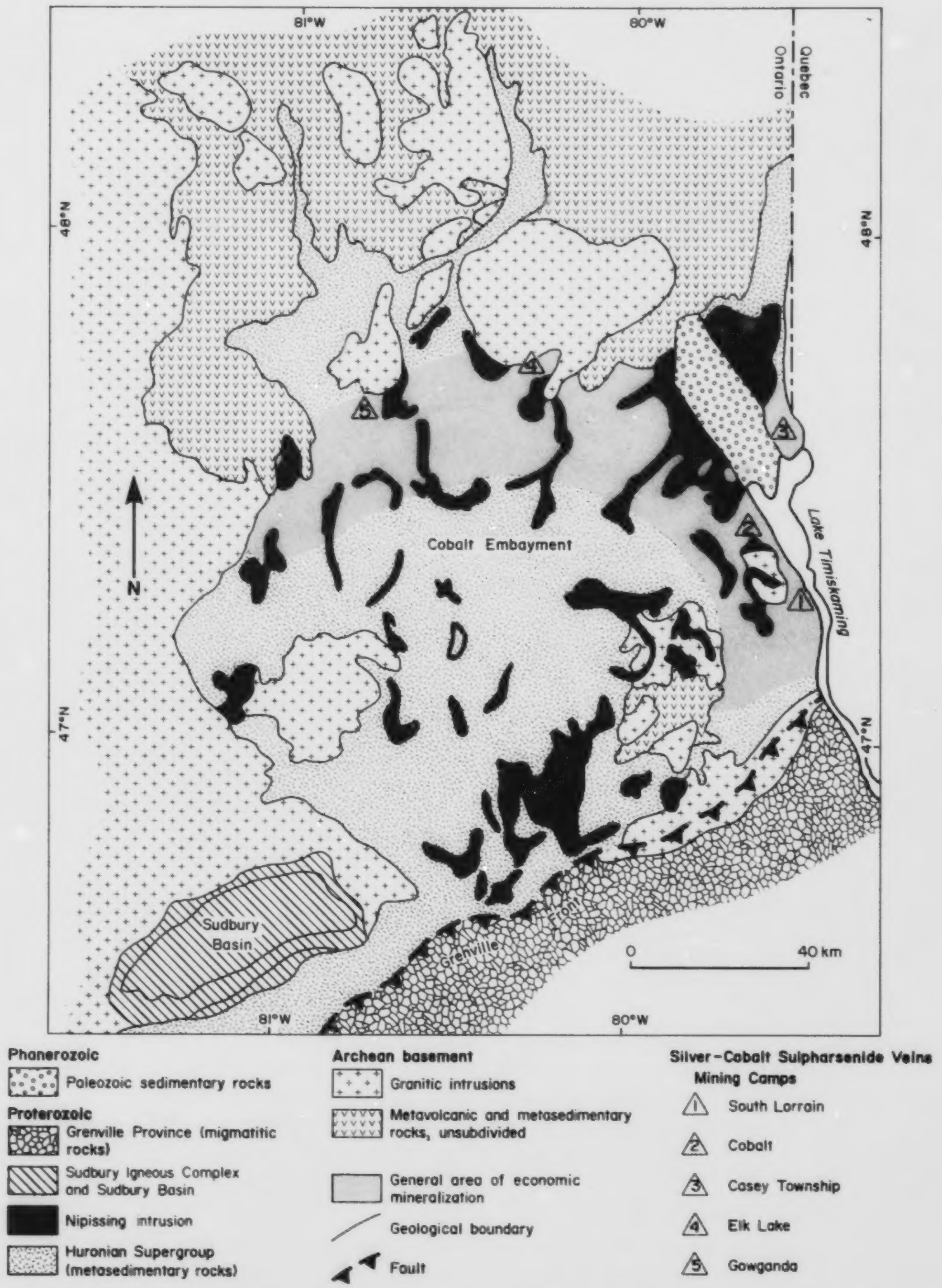

81°W
80°W
Quebec
Ontario
48°N
48°N
47°N
47°N
N
Cobalt Embayment
Lake Timiskaming
Sudbury Basin
Grenville Front
0
40 km
81°W
80°W
Phanerozoic
Paleozoic sedimentary rocks
Proterozoic
Grenville Province (migmatitic rocks)
Sudbury Igneous Complex and Sudbury Basin
Nipissing intrusion
Huronian Supergroup (metasedimentary rocks)
Archean basement
Granitic intrusions
Metavolcanic and metasedimentary rocks, unsubdivided
General area of economic mineralization
Geological boundary
Fault
Silver–Cobalt Sulpharsenide Veins
Mining Camps
1 South Lorrain
2 Cobalt
3 Casey Township
4 Elk Lake
5 Gowganda

**Figure 23.9.** **a)** Distribution of silver-cobalt sulpharsenide and amethyst veins in the Thunder Bay area and **b)** silver-cobalt sulpharsenide veins within the Cobalt Embayment. *Modified from* Andrews et al. (1986) and Franklin et al. (1986).

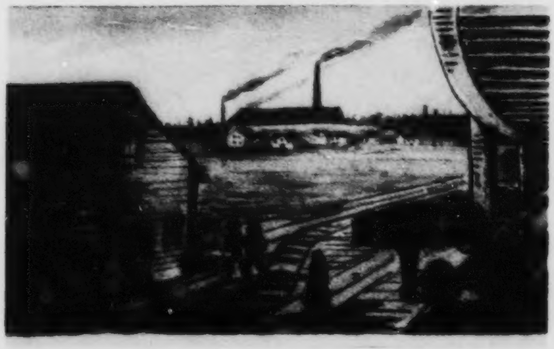


**Figure 23.10.** View towards the mill of the West Canada Mining Company, Bruce Mines. Bruce Mines was the first commercial copper mining establishment in Canada. Mining operations were active from 1847–1875, and again between 1898 and 1921 (painting by Elliot, displayed in the Archives of Ontario; photograph *courtesy of* E.J. Leahy, MNDM Sault Ste. Marie).

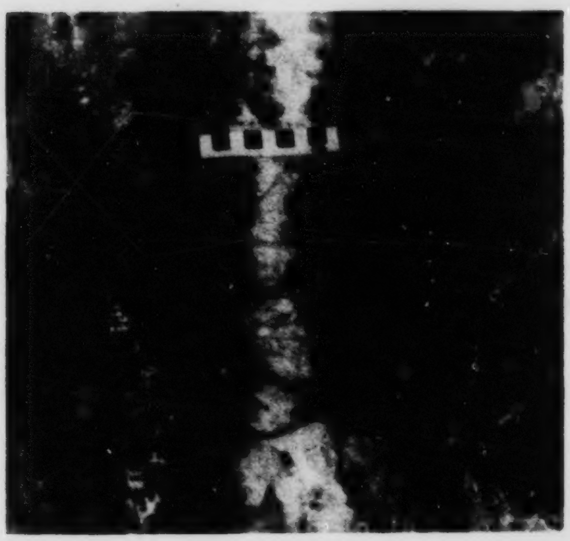


**Figure 23.11.** Zoned quartz-carbonate-specularite-chalcopyrite vein, near Iron Bridge, Gladstone Township. Vein cuts siltstone of the Gowganda Formation. Dark edges of vein consist of iron-bearing carbonate. Specularite and chalcopyrite occur in the central, quartz-rich part of the vein (photograph *courtesy of* G. Bennett, MNDM, Sault Ste. Marie).

Keweenawan faults is consistent with vein formation during and after Keweenawan rifting (Franklin 1970).

## Gold, Quartz, Arsenopyrite, Pyrite and Chalcopyrite Veins

Within the Huronian Supergroup, most veins having this mineral assemblage are concentrated south of Espanola; a few also occur southeast of Wanapitei Lake (*see* Figure 23.8). Although these veins occur most commonly in rocks of the Gowganda Formation and Nipissing gabbro, veins also occur in sedimentary rocks of the Mississagi, Espanola and Serpent formations (Pearson 1980; Gates 1991). The veins consist of native gold, pyrite, arsenopyrite, chalcopyrite and pyrrhotite (Pearson 1980). There are also several gold-bearing quartz-arsenopyrite-pyrite vein occurrences in the Espanola area. Of these, the Bousquet Mine produced 145 299 g of gold between 1936 and 1938 (Card 1976).

## Silver-Sulpharsenide Veins

In Ontario, occurrences and deposits of this type occur in the Thunder Bay area (Franklin et al. 1986; Kissin and Jennings 1987; Kissin and Sherlock 1989) and along the north and northeastern margins of the Cobalt Embayment (*see* Figure 9b) in the Cobalt and Gowganda camps (cf., Berry 1971; Patterson 1979; Appleyard 1981; Legun 1981; Owsiacki 1981, 1982, 1984; Donaldson et al. 1984). The deposits in the Huronian Supergroup have yielded nearly 500 million ounces of silver (Owsiacki 1984), whereas those in the Thunder Bay area have yielded about 4.7 million ounces of silver (Sergiades 1968). The sulpharsenide association is characterized by: 1) the occurrence of silver as the principal economic metal, predominantly in the native state; 2) the association of silver with arsenides and sulpharsenides of cobalt, nickel, iron and antimony, which usually occur as the main metallic components of the veins; and 3) the presence of calcite and dolomite as dominant gangue minerals.

### THUNDER BAY AREA

Silver-bearing veins in the Thunder Bay area (*see* Figure 23.9a) cut the Rove and Gunflint formations and gabbro dikes, although most occur within rocks of the Rove Formation (Franklin 1970). In general, the veins are complexly zoned. Silver-poor vein segments consist of calcite, quartz, fluorite, minor barite and less common sulphide minerals (Franklin et al. 1986). Silver-rich vein segments are banded and can consist of a pyrite-rich core surrounded by calcite and quartz (Franklin et al. 1986). The deposits are divided into 3 arrays (Oja 1967; Franklin 1970; Franklin et al. 1986): 1) the Mainland vein array, that occurs along northeast-striking normal faults near the western margin of the Proterozoic rocks; 2) the Island vein array, that occurs in, or near to, a northeast-striking swarm of gabbro dikes, but within northwest-striking fractures; and 3) veins, that occur on the western side of the Port Coldwell alkalic complex. The Port Coldwell vein array cuts Archean metasedimentary rocks and Keweenawan gabbroic dikes (Franklin et al. 1986). These veins have not been comprehensively studied and are not discussed in detail.

The veins in the Mainland array (*see* Figure 23.9a) occur within a northeast-striking fault zone that forms part of the boundary zone that separates Archean from Proterozoic rocks. Most of the veins cut the Rove and Gunflint formations, and only a few veins are entirely contained within a diabasic Logan sill (Franklin et al. 1986). The veins in both the Logan sills and the Gunflint Formation do not contain as much silver as those in the Rove Formation (Franklin et al. 1986). The veins or their wall rocks are generally not deformed (Franklin et al. 1986). Alteration adjacent to the veins consists typically of silicification of the Rove Formation and hydration of the diabase (Franklin et al.

1986). The veins strike approximately 065° parallel to the strike of major faults (Franklin et al. 1986). This vein set contains acanthite, pyrite, sphalerite, marcasite, galena, calcite, quartz, barite and fluorite (Franklin et al. 1986).

The Island vein array (*see* Figure 23.9a) includes the productive Silver Islet deposit. The veins in this array occur within the upper Rove Formation. The most productive parts of the veins are in gabbroic dikes that strike 060°, parallel to the northwestern shore of Lake Superior (Franklin et al. 1986). The veins strike 340° perpendicular to the strike of the gabbro dikes (Franklin et al. 1986). Veins in the Rove Formation generally have lower silver and higher base metal tenors than veins which occur within the central parts of the gabbro dikes (Franklin et al. 1986). Wall rock alteration consists of replacement by laumontite in the metasedimentary rocks and hydration of the gabbro dikes (Franklin et al. 1986). Based on vein mineralogy, the Island vein array is subdivided into a silver-cobalt-nickel-arsenic-bismuth sulpharsenide type and a base metal sulphide-barite type with variable silver content (Oja 1967; Smyk 1984). Vein constituents include native silver, acanthite, arsenides, pyrite, marcasite, galena, sphalerite, chalcopyrite, quartz, dolomite, calcite, barite and fluorite (Franklin 1970; Smyk 1984; Franklin et al. 1986). The mineral assemblages of the 2 vein types include (Franklin 1970; Smyk 1984; Franklin et al. 1986): 1) calcite, barite, galena, sphalerite and chalcopyrite for the base metal sulphide-barite type; and 2) sulfosalt and sulphide minerals with native silver, dolomite, quartz, barite and fluorite for the nickel-cobalt-silver sulpharsenide type.

The close spatial association between the veins and the contacts of the Rove Formation, with either diabasic Logan sills or northeast-striking gabbro dikes, indicates a depositional relationship. The dikes are inferred to have formed structural traps for hydrothermal fluids (Franklin 1970; Franklin et al. 1986). The veins formed in Keweenawan faults related to intracontinental rifting. These faults were active after the principal Keweenawan volcanism; thus, the vein formation postdates the 1.11 billion-year-old Logan sills and possibly postdates the accumulation of volcanic and sedimentary rocks of the 1.09 billion-year-old (Davis and Sutcliffe 1985) Osler Formation.

Fluid inclusion data indicate that the hydrothermal fluids were heated at depth as they ascended, cooled from in excess of 370° to 100°C, boiled, underwent dilution and precipitated minerals at about 1 km depth in structurally favourable sites beneath diabase sills (Kissin and Sherlock 1989). Mixing between a saline, hot brine and dilute meteoric water may account for variable fluid inclusion salinities (Kissin and Sherlock 1989). Heat associated with Keweenawan rifting is inferred to have mobilized connate brines, which scavenged metals at high temperatures (Kissin and Sherlock 1989). Periodic influx of hot brines, followed by fluid cooling and deposition of vein minerals, is attributed to periodic fault movement and repeated vein opening (Kissin and Sherlock 1989).

## COBALT CAMP—HURONIAN SUPERGROUP

Silver-sulpharsenide veins (*see* Figure 23.9b) occur mainly in steeply dipping, anastomosing arrays that range from less than a centimetre to more than 20 cm in width. The vein systems are mostly fault-controlled with mineralization occurring principally within the Coleman Member of the Gowganda Formation, Nipissing gabbro and, to a lesser extent, Archean supracrustal rocks (cf., Andrews et al. 1986). Many of the faults that crosscut the diabase are filled with silver-bearing veins (Whitehead 1920; Knight 1924; Moore 1956; Hester 1967; McIlwaine 1970). All the economically productive deposits occur in close proximity to the Huronian-Archean unconformity where Nipissing gabbro and steeply dipping Archean volcanic sequences coincide (Andrews et al. 1986). A spatial association exists between base metal sulphide mineralization in the Archean basement and the cobalt-silver-arsenic veins that cut the unconformably overlying Huronian Supergroup (Patterson 1979). Silver occurrences are known, but are relatively uncommon, in rocks of the Firstbrook Member of the upper Gowganda Formation and the Lorrain Formation in the upper part of the Huronian sequence. These rocks dominate in the central part of the Cobalt Embayment. No deposits of economic significance are known to occur above, and remote from, the Archean-Huronian unconformity (cf., Andrews et al. 1986).

All economic deposits occur spatially associated with Nipissing gabbro. The deposits occur either within the Nipissing gabbro or within 200 m of its upper and lower contacts (cf., Andrews et al. 1986). In contrast to the localization of mineralized zones, the vein systems can be quite extensive and, in some cases, cut completely across the intrusions of Nipissing gabbro and commonly continue for significant distances into surrounding country rocks (cf., Andrews et al. 1986).

Vein formation involved the precipitation of mainly quartz, chlorite, actinolite and potassium feldspar during initial, limited dilation, followed by the introduction of significant quantities of calcite and dolomite (Jambor 1971a) during subsequent dilational episodes. Carbonates compose the dominant gangue component and typically occupy the main, central part of the veins (Andrews et al. 1986). Silicate minerals are limited to thin (less than 1 cm) selvages immediately adjacent to vein walls (Andrews et al. 1986). The silver-bearing mineral assemblage, when present, occurs at, or near, the transition from the silicate- to the carbonate-bearing assemblage and, therefore, is preferentially distributed immediately adjacent to vein walls (Andrews et al. 1986).

The metallic mineralogy is complex, including native silver and bismuth, arsenides, sulphides and minor oxides (Petruk 1971). Arsenide minerals are volumetrically the most important and can be classified into nickel-arsenic, nickel-cobalt-arsenic, cobalt-arsenic, cobalt-iron-arsenic and iron-arsenic assemblages with the cobalt-arsenic types being most abundant (Petruk 1971). Sulphide minerals consist primarily of chalcopyrite, tetrahedrite, galena, sphalerite, marcasite and pyrite (Boyle 1968; Petruk 1971;

Petruk, Harris, Cabri and Stewart 1971; Petruk, Harris and Stewart 1971; Petruk and Staff 1971; Jambor 1971a). Metallic minerals occur in a wide variety of forms including massive pods, bands, dendrites, plates, leaves and zoned rosettes (Figure 23.12; cf., Andrews et al. 1986). Silicate and carbonate vein mineralogy appears to be independent of host rock type or deposit environment and does not vary systematically between barren and mineralized zones (Andrews et al. 1986). Mineral precipitation within the veins consistently follows the sequence of early silicate formation, followed by ore minerals, dolomite, where present, and finally by the deposition of relatively large quantities of calcite (Andrews et al. 1986).

The mineralization is typically discontinuous along any given vein structure. High-grade ore pockets commonly occur in the vicinity of vein intersections; intersections of veins with late shallow-dipping shear zones; rock contacts; and abrupt changes in the configuration of the Archean basement topography that may be fault controlled (Andrews et al. 1986). Mineral zonation of silver and iron, cobalt, nickel arsenides has been documented within individual vein systems (Petruk 1971; Scott 1972; Robinson 1984) but is locally variable. Some veins formed within simple fissure openings and in zones of late dilation along pre-existing faults and shear zones (Robinson 1984; Andrews et al. 1986). Other veins have been deformed, characterized by severe grain-size reduction of vein minerals, penetrative fabric development parallel to vein walls and drag folding of massive arsenide bands (Andrews et al. 1986).

Subhorizontal vein arrays, which occupy mainly shallow-dipping (less than 20°) faults and shear zones, also occur. These veins are dominated by quartz, potassium feldspar and epidote with minor carbonate, chalcopyrite, sphalerite, galena, axinite and occasionally bismuth (Andrews et al. 1986). They are generally devoid of silver and sulpharsenide mineralization except where they intersect with steeply dipping, silver-bearing vein structures (Andrews et al. 1986). Although equivocal, the shallow-dipping veins appear to have formed contemporaneously with, to slightly later than, the silver-sulpharsenide mineralization (Andrews et al. 1986).

Wall rock alteration haloes exhibit a similar silicate to carbonate paragenesis as is evident in the veins. Secondary albite commonly occurs in wall rocks adjacent to Nipissing gabbro, and sporadically distributed potassium feldspar can occur in Archean basalt wall rocks (Andrews et al. 1986). Alteration assemblages are consistent with the introduction of $CO_2$, calcium, sodium, potassium, silver, arsenic, cobalt, lead, rare earth elements (REEs) and, in some cases, mercury and gold by hydrothermal fluids (Andrews et al. 1986). This was accompanied by significant net loss of silicon, iron, magnesium, zinc, boron, lithium and scandium from the wall rocks (Andrews et al. 1986).

## ORIGIN OF SILVER-COBALT-ARSENIC VEINS

The origin of the veins has been debated since their discovery (Knight 1924; Boyle 1968; Jambor 1971b; Andrews et al. 1986). Boyle (1968) and Patterson (1979) stressed the importance of an Archean source for the metals. Boyle (1968) concluded that the Nipissing gabbro may have contributed nickel and cobalt and the heat required to drive a hydrothermal system.

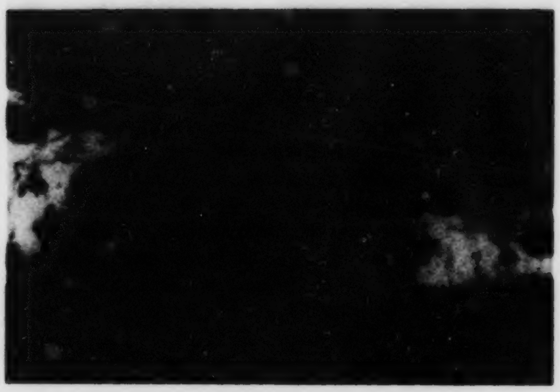


**Figure 23.12.** Cobalt-nickel-arsenide-silver-bearing calcite vein, Silverfields Mine, Cobalt. Core of "obicular" form consists of niccolite (NiAs), that is surrounded by breithauptite (NiSb) and native silver. Location-1337-south stope; vein thickness is 15 cm (photograph *courtesy of* J. Ireland, MNDM, Cobalt).

Several considerations discourage the theory that the Nipissing gabbro acted purely as a heat and/or fluid source in vein formation. Andrews et al. (1986) demonstrated that wall rock alteration adjacent to the veins postdated a low temperature, regional alteration assemblage that had affected the intrusions and chlorite spotting alteration in the Huronian metasedimentary rocks. The chlorite spotting is attributed to contact metamorphism accompanying the intrusion of Nipissing gabbro (Andrews et al. 1986). These data are consistent with vein formation after the emplacement of the Nipissing gabbro and much (possibly all) of the Nipissing gabbro cooling history (Boyle 1968; Andrews et al. 1986). Furthermore, intrusions of Nipissing gabbro are widely distributed throughout, and beyond, the Cobalt Embayment, but the vein deposits are restricted to the northern and northeastern boundaries of this broad domain (*see* Figure 23.9b). Finally, the spatial association between silver-sulpharsenide deposits and Nipissing gabbro evident in the Cobalt Embayment is, on a global basis, not typical (Bastin 1939; Badham 1976).

Andrews et al. (1986) and Franklin et al. (1986) proposed that the Nipissing gabbro behaved as mechanically favourable sites for fracture generation during regional fault activity. There is evidence that the Cobalt Embayment experienced repeated tectonic activity before, during and after the intrusion of Nipissing gabbro, resulting in the generation and reactivation of regional-scale faults (cf., Andrews et al. 1986). The flat-lying sills could represent mechanical-geometrical anisotropies favourable to fracture generation where steeply dipping faults intersect them. This would apply particularly where the flat-lying sills occur within, or in close proximity to, the steeply dipping Archean basement volcanic rocks and less so where they occur within the flat-lying Huronian sequence (Andrews et al. 1986). This rationale conforms with the observation that all the mined deposits occur in proximity to the Archean-Huronian unconformity and specifically where the Nipissing gabbro has been emplaced within, or immediately adjacent to, steeply dipping Archean volcanic rocks (Andrews et al. 1986). Whereas, occurrences of silver-sulpharsenide mineralization are found throughout the Huronian sequence (always within or adjacent to sills), no deposits of economic significance are known to occur remote from the Archean unconformity. Furthermore, the spatial association between base metal sulphide mineralization in the Archean basement and the cobalt-silver-arsenic veins is consistent with a metal source for the Archean mineralization (Patterson 1979).

Veins in the Thunder Bay area are localized in faults that were active during 1.1 billion-year-old Keweenawan volcanism and sedimentation (Franklin 1970). However, most veins were formed late in this tectonic history, as few veins are deformed (Franklin 1970).

## Thunder Bay Amethyst Deposits

Amethyst is a semiprecious gemstone variety of quartz (Figure 23.13). Although clear or white varieties of quartz are quite common, some of the coloured varieties, such as amethyst, are not. There are only 2 known Canadian localities which have continuously produced gem quality amethyst, one of those is the area north of Lake Superior, near Thunder Bay, the other is Nova Scotia. Increasing tourist interest in the Thunder Bay area, unique in the quantity of high-quality amethyst deposits, prompted the Government of Ontario to name amethyst as the Provincial Mineral Emblem on May 14, 1975.

Amethyst mining and related tourist operations constitute a local industry in the Thunder Bay area (*see* Figure 23.9). The amethyst mines operate seasonally as tourist sites, as well as catering to customers on an on-demand basis. Prospectors with mining leases or quarry permits also supply customers. An estimated 1991 production value, based on an average of $1.00 per pound, is approximately $500 000.00.

### MINERALOGICAL AND COLOUR CHARACTERISTICS

Amethyst is the purple variety of alpha quartz, which is the lower temperature version of beta quartz. The structure of alpha quartz is trigonal. The crystals are commonly prismatic and are terminated by 2 rhombohedral sets.

### AMETHYST COLOUR

Amethyst colour varies from light mauve or violet to almost black. The purple colour results from a complex substitutional and ionization process. Alpha quartz accommodates interstitial ferric iron (Cohen 1985). Aluminum is also present in the alpha quartz but as a substitute for silicon in the tetrahedral structure. The charge on the aluminum is compensated by an interstitial cation, such as sodium. When subjected to ionizing radiation at low temperatures, the bonded aluminum-oxygen pair loses an electron, and the resultant configuration is referred to as a "trapped-hole" (Cohen 1989). The electron is captured by the interstitial sodium. Cohen (1985, 1989) and Adekeye and Cohen (1986) demonstrated that ionizing radiation causes interstitial ferric iron to lose an electron. The electron from this reaction is recaptured by the aluminum-oxygen pair to neutralize the trapped-hole. The colour of amethyst is attributed to the interstitial $Fe^{4+}$ ion (Cohen and Hassan 1974). The quenching of the trapped-hole suppresses the formation of smoky quartz.

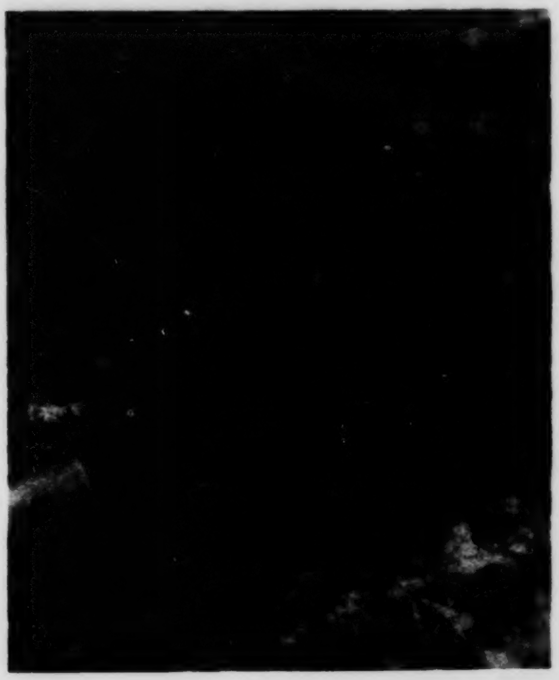


Figure 23.13. Vug at the Diamon Willow Mine, owned by Doug Noyes (photograph *courtesy of* M.I. Garland, MNDM Thunder Bay).

The interstitial ferric iron ion causes some strain in the quartz crystal lattice, resulting in the formation of quartz twinned according to the Brazil Law. Amethyst colour is only seen in quartz that is twinned according to the Brazil Law (Cohen 1985; Adekeye and Cohen 1986). Heating amethyst to temperatures of 350°C or higher reverses the colour process; the quartz changes from purple to yellow and then becomes colourless (Cohen and Hassan 1974; McLean 1984).

### GROWTH CONDITIONS

The formation of amethyst requires precipitation of quartz from hydrothermal solutions containing ferric iron in a near-surface oxidizing environment. It is also necessary for the hydrothermal solutions to stay within the temperature range of alpha quartz and to ensure that ferric rather than ferrous iron dominates. The formation of amethyst also requires a source of ionizing radiation, necessary to induce the trapped-holes, and a system of open fissures or cavities must be maintained so that the quartz crystallizes in a crystalline, rather than massive form.

## AMETHYST DEPOSITS IN THE THUNDER BAY AREA

Amethyst veins in the Thunder Bay area (*see* Figure 23.9) occur in parallel to subparallel vein systems, rather than anastomosing quartz veins (Garland 1992). Veins strike 045°, parallel to major Keweenawan faults, or 090°±10° (Garland 1992). The best amethyst crystals occur in vugs and the largest vugs occur at the intersection between the 045° and 090° faults and fractures. Because amethyst formation is restricted to near-surface conditions, all amethyst occurrences are located at, or near, the subhorizontal unconformity between the Archean and Proterozoic rocks. The unconformity has also acted as a mechanism for concentrating radioactive minerals essential in the formation of the purple colour of amethyst.

### DEPOSIT CHARACTERISTICS

Amethyst veins cut Archean granite or monzonite, Proterozoic mudstone or conglomerate units of the Proterozoic Sibley Group or the Rove and Gunflint formations. There is usually a fault and/or hydraulic breccia associated with the quartz veins (Garland 1992). The breccia zones and fragments are developed on a metre scale and any of the host rocks can be present as fragments. Some breccias, developed in Archean host rocks, consist only of Proterozoic rock fragments. It is also possible to have multiple generations of brecciation. The fragments in the breccia are cemented together by white quartz.

Amethyst occurs as one of the later generations of quartz deposition, forming a crystal druze filling angular interfragment spaces (Garland 1992). If there are no spaces, then amethyst is not present. Amethyst also crystallizes along the edge of the breccia zones in relatively narrow, but extensive veins that form large walls of spectacular crystal druze. One such breccia zone, which can be traced along strike for a kilometre, is present at the Diamond Willow and Blue Points Amethyst mines in McTavish Township (*see* Figure 23.9a). Breccia zones are present at all the producing amethyst mines in McTavish Township and at all the known occurrences of amethyst in that area (Garland 1992).

The amethyst veins in MacGregor Township (*see* Figure 23.9a) occur in an Archean granite and strike parallel to the preferred 045° and 090° directions. The lack of any well developed breccia zones distinguishes these amethyst veins from those in McTavish Township. The veins consist of networks of white quartz stringers in pink, texturally massive granite (Garland 1992).

Where the openings are in the form of fractures, the amethyst tends to form massive, purple veins, with ghosts of crystal points outlined in the quartz. These veins can be up to 30 or 40 cm in thickness. Where the openings are vugs, the amethyst lines the vugs in magnificent crystal druze, with points (crystal terminations) up to 15 cm in diameter. Vugs occur for no apparent reason along fractures and at the intersection between the faults and fractures which have a strike of 045° and 090°. Vug size is variable but can range up to 4 m in the vertical dimension. Size of the amethyst points is directly related to the size of the vug.

Offset along the fault zones is normal. In the McTavish Township cluster (*see* Figure 23.9a), at the Ontario Gem Mine and the Diamond Willow and Blue Points mines (southeast of the Pearl Lake Mine), the north side of the fault has moved down relative to the south side (Garland 1992). At the Pearl Lake Mine, the south side of the fault has moved down apparently defining a fault block. Lateral movement along the amethyst-containing faults has not been determined due to lack of crosscutting features (Garland 1992). Movement of any kind along the faults within Archean rocks is difficult to determine because of the absence of markers.

### VEIN MINERALOGY

The mineralogy of amethyst occurrences varies regionally. West of Thunder Bay, amethyst is spatially associated with the silver-calcite veins and occurs as a minor constituent within vein systems containing calcite, barite, galena, sphalerite, fluorite and silver minerals. East of Thunder Bay, the amethyst vein system is dominated by quartz accompanied by minor amounts of barite, calcite, galena, sphalerite and minor chalcopyrite. These amethyst veins are spatially associated with the calcite, galena, sphalerite and silver veins (Garland 1992).

There is mineral zonation within the amethyst veins, usually due to the presence of several generations of quartz. Early formed quartz is generally white. If barite is present, it occurs as orange rosettes on amethyst, which grew at the outer edges of the zone. Calcite is usually restricted to irregular patches within the vein system and also occurs near the outer edges. Galena, if present, occurs as blebs within the calcite. Zoning within the amethyst crystals is exhibited by changes in crystal colour and the temperature and composition characteristics of trapped fluid inclusions (McArthur and Kissin 1988).

### PRECIPITATION CONDITIONS

Two stages of precipitation, separated by a period of fluid cooling, are documented in the amethyst vein systems (McArthur and Kissin 1988). Smoky quartz, followed by amethyst, precipitated at approximately 140°C and 60° to 90°C, respectively (McArthur and Kissin 1988). Trapped fluid inclusions contain about 23 to 15 weight percent $NaCl_{equivalent}$; salinity of these fluids decreases in the direction of quartz crystal growth (McArthur and Kissin 1988). Minor pyrite and chalcopyrite, contained in the amethyst veins, have $\delta^{34}S$ values of -0.4 to -1.4‰ similar to lead-zinc-barite veins in the area (McArthur and Kissin 1988). The amethyst veins are interpreted to have formed from evolved basin brines, expelled from the Sibley Group rocks, along, or parallel to, Keweenawan faults (McArthur and Kissin, 1988).

## BRECCIA- AND INTRUSION-ASSOCIATED COPPER MINERALIZATION

Keweenawan breccia pipe- and intrusion-hosted disseminated copper mineralization occur in the Batchawana area (*see* Figure 23.3) of southeastern Lake Superior (Blecha 1974; Pearson 1980). The Breton, West, South and East breccia pipes collectively compose the Tribag breccia cluster which occurs, at the western end of the Archean Batchawana greenstone belt, near the contact between the greenstone belt and Archean granites to the north (Blecha 1974). The Palmer breccia pipe and the Jogran intrusion-hosted, disseminated copper deposit also occur in the Batchawana greenstone belt, south and southwest of the Tribag cluster respectively (Blecha 1974). Higher grade mineralized zones were mined from the Breton Pipe in the Tribag cluster (Blecha 1974).

Fragment types within the breccia pipes consist of granite, diabase, basalt and felsite set in a quartz-calcite matrix (Figure 23.14; Blecha 1974). The proportion of matrix and fragment type varies between pipes (Blecha 1974). Metallic minerals include chalcopyrite, pyrite, pyrrhotite, molybdenite, minor scheelite, sphalerite, galena and rare wolframite, stibnite and native copper (Armbrust 1969; Blecha 1974). These minerals can be disseminated uniformly throughout the matrix or rock fragments, along late fractures, or concentrated in localized zones within a pipe (Blecha 1974). Hydrothermal quartz, 1.05 billion-year-old (Roscoe 1965) sericite and kaolinite are abundant in the mineralized areas of the Breton Pipe (Armbrust 1969).

The Jogran copper deposit consists of disseminated and fracture-filling pyrite, chalcopyrite, minor amounts of molybdenite and rare galena within a quartz-feldspar porphyry (Armbrust 1979). The porphyry body was intruded into Archean mafic metavolcanic rocks (Giblin 1966; Blecha 1974). Potassic alteration is manifested by quartz and potassium feldspar in the porphyry and biotite in the adjacent basalt (Armbrust 1979). Quartz-sericite phyllic and albite-epidote-chlorite-calcite-hematite propylitic alteration are also developed (Armbrust 1979). Higher copper tenor is associated with the potassic alteration zone (Shklanka 1969, p.60).

All of these deposits are considered to be of porphyry type, their pipe-like shape suggesting derivation from a deep seated felsic magma (Blecha 1974). While the Jogran copper deposit displays characteristics of porphyry-type mineralization (Armbrust 1979), the breccia pipes may also have formed by diatreme activity associated with Keweenawan felsic magmatism (Pearson 1980). Associated hydrothermal fluids were of magmatic origin (Norman 1977).

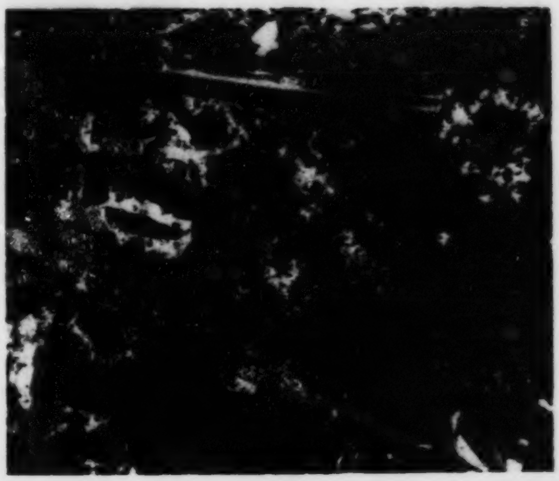


**Figure 23.14.** Tribag-type breccia—West Breccia Pipe, located 1 km south of the Tribag Mine (Breton Breccia pipe), Nicolet Township. Breccia fragments consist of angular to subrounded Archean mafic metavolcanic rocks, and minor Keweenawan felsite and diabase. Chalcopyrite, scheelite and minor wolframite occurred in cementing quartz (photograph *courtesy of* G. Bennett, MNDM, Sault Ste. Marie).

### Regional Sodium Metasomatism—Huronian Supergroup

Sodium metasomatism, although centred on the Sudbury Igneous Complex (Meyer et al. 1990; Gates 1991), is widespread in the Huronian Supergroup between Sault Ste. Marie and Cobalt (*see* Figure 23.8; Roscoe 1969; Meyer et al. 1986, 1987). This secondary albite is associated with many of the described metallic occurrences and deposits (Meyer et al. 1986, 1987). Most notable is the association with the Scadding Gold Mine (Orofino Resources Limited) in Scadding Township and the Norstar Mine (Groundstar Resources Limited-Orofino Resources Limited) copper-gold deposit in Davis Township, located east of the Sudbury Basin (*see* Figure 23.8). Because of its widespread nature and the association with many of the metallic concentrations, a brief discussion of the sodium metasomatism is warranted.

The sodic alteration is characterized by replacement of wall rock by albite (Figure 23.15; Meyer et al. 1987). The alteration is manifested by veins, dike-like zones, stratabound, areally extensive massive zones and zones of brecciated, altered rock (Meyer et al. 1987). In some albitized zones at the Scadding gold, Crystal and Norstar mines, secondary albite is commonly replaced by carbonates, which are, in turn, replaced by quartz and chlorite (Meyer et al. 1987, 1989; Gates 1991). Gold in these settings occurs in quartz-carbonate veins or quartz-chlorite zones that cut or replace albitized rock (Meyer et al. 1989; Gates 1991). Pyrite, pyrrhotite, chalcopyrite, galena, gold and arsenopyrite also occur in veins or as disseminated zones that cut the zones of albitized rock (Figure 23.16; Meyer et al. 1989, 1990; Gates 1991).

Whereas most gold and copper-gold vein deposits and occurrences in the area of Wanapitei Lake appear to be related spatially and temporally with zones of albitized country rock, there is also some mineralization that is spatially and temporally associated with secondary chlorite-rich assemblages that replace the albitized rock (cf., Meyer

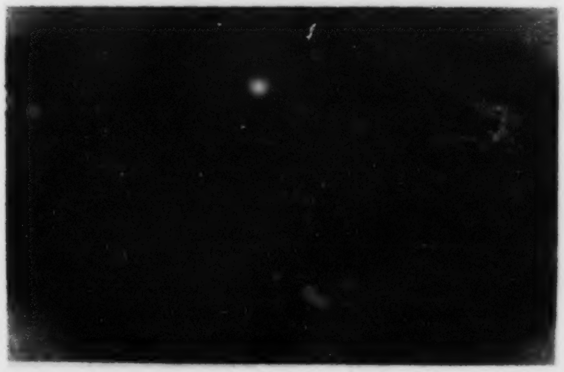


Figure 23.15. Albitized Gowganda Formation siltstone from a gold occurrence in Davis Township. Light material is albite, dark rhombohedral crystals are iron- and magnesium-bearing carbonates, some of which have been replaced by pyrite (photograph *courtesy of* B. Gates, MNDM, Sudbury).

Figure 23.16. Albitized Mississagi Formation quartzite. Light coloured material to right of coin is hematite-bearing albite, granular material is pyrite that contains up to 31 g/t Au, 0.2 wt% Co and 0.2 wt% Ni (Gates 1991). The alteration and pyrite mineralization is developed along fractures and as zones up to 1 m in thickness (photograph *courtesy of* B. Gates, MNDM, Sudbury).

et al. 1988, 1989, 1990; Gates 1991). Therefore, at least some gold and copper-gold mineralization was introduced during both the albitization and chloritization events.

Albitized rocks include Huronian metasedimentary rock, 2.22 billion-year-old (Corfu and Andrews 1986) Nipissing gabbro, the matrix of Sudbury breccias and pseudotachylite related to the 1.85 billion-year-old (Krogh et al. 1984) Sudbury Event (Meyer et al. 1987). Many albitized zones, spatially associated with copper-gold mineral occurrences, are also spatially associated with zones of Sudbury Breccia (Meyer et al. 1990). However, west- to northwest-striking 1.24 billion-year-old Sudbury diabase dikes are not albitized (Meyer et al. 1987). Hydrothermal monazite crystallized at approximately 1.7 Ga (Schandl et al. 1991) during the sodium metasomatism of Huronian metasedimentary rocks south of Wanapitei Lake. Although the duration and number of discrete sodium metasomatic events remains poorly documented, the 1.7 billion-year-old metasomatic event correlates with a proposed 1.7 billion-year-old orogeny that affected the northwestern part of the Grenville Province in Ontario (Easton 1992). This 1.7 billion-year-old orogeny may have involved the development of the 1.74 billion-year-old Killarney Magmatic Arc, the formation of 1.74 and 1.69 billion-year-old sedimentary sequences and high-grade metamorphism at 1.68 Ga, possibly due to crustal stacking during continental collision (Easton 1992).

Thus, the sodium metasomatism and spatially associated gold and copper-gold occurrences and deposits formed during a significant and widespread event that affected sedimentary and igneous Proterozoic rocks in the area east of Lake Huron (Meyer et al. 1986, 1987, 1988, 1989, 1990). This event may have been a circa 1.7 billion-year-old orogeny that affected the northwestern part of the Grenville Province in Ontario (Easton 1992).

## Regional Metal Distribution Patterns—Huronian Supergroup

Regional metal distribution patterns are developed in the Huronian Supergroup (Card and Pattison 1973; Innes and Colvine 1979, 1984; Figure 23.17). Metal distribution patterns that are developed for the hydrothermal veins include (Innes and Colvine 1979, 1984):

1. With increasing distance from Sudbury, vein-type mineralization, hosted preferentially within Nipissing gabbro, becomes more abundant, and vein gangue changes from dominantly quartz to dominantly carbonate.
2. Copper-bearing, chalcopyrite and bornite veins are more abundant in the western part of the area.
3. Polymetallic, base metal veins are more frequent in the northeastern part the area.
4. Silver-cobalt sulpharsenide veins predominate in the extreme northeastern part of the area.
5. Gold-bearing veins predominate south of Wanapitei Lake and in the Sault Ste. Marie areas.

Sulphide mineralization associated with Nipissing gabbro also shows metal distribution patterns. In the Sudbury area, sulphide mineralization associated with Nipissing gabbro is characterized by the presence of disseminated pyrrhotite, chalcopyrite and pentlandite with accessory PGE elements within the gabbro dikes (Innes and Colvine 1984). With increasing distance from Sudbury, the abundance of pentlandite, total sulphur and PGE tenor decrease, and nickeliferous pyrrhotite becomes the nickel host (Innes and Colvine 1984). Metal zoning of the magmatic nickel-copper-PGE sulphide mineralization appears to correlate with geometric aspects of Nipissing gabbro. Nipissing gabbro, proximal to Sudbury, dip steeply, but dip more gently and have an increased volume of granophyric phase as the distance from Sudbury increases (Card and Pattison 1973). The coincidence of higher pentlandite and total sulphur abundance and nickel and PGE tenors, with the dip relationships of Nipissing gabbro, may indicate that Sudbury was a locus of 2.2 billion-year-old magmatism (Innes and Colvine 1984).

These regional patterns have been interpreted to reflect:

1. tapping of a metal and sulphur-rich mantle region during generation of Nipissing mafic magma (Card and Hutchinson 1972; Card and Pattison 1973; Innes and Colvine 1979)
2. sulphur saturation of Nipissing mafic magma induced by assimilation of sulphur-bearing, lower Huronian rocks (Pearson 1980)
3. redistribution, into hydrothermal veins, of metals previously concentrated in Huronian Supergroup sedimentary rocks. This redistribution occurred, by paleoplacer processes, during the emplacement of Nipissing gabbro (Innes and Colvine 1979, 1984).
4. exposure of a deeper crustal section, in the Sudbury area, associated with a possible centre or focus of Nipissing gabbro emplacement (Innes and Colvine 1984).

A genetic relationship between Nipissing gabbro and associated magmatic nickel-copper-PGE sulphide mineralization is likely. However, presently, there is no evidence to support the existence of more than one Nipissing magma type (P. Lightfoot, Ontario Geological Survey, personal communication, 1990). Thus, the metal distribution patterns cannot be simply attributed to the involvement of fundamentally different Nipissing magma types. However, the composition of the Nipissing magma type is consistent with the assimilation of a crustal component (Lightfoot and Naldrett 1989). Such assimilation would have been widespread and consistent, given the regionally consistent and uniform nature of the parental Nipissing magma (P. Lightfoot, Ontario Geological Survey, personal communication, 1990). If crustal assimilation had taken place in a centralized magma source region, the resultant sulphide accumulations might be expected to have been deposited in proximity to magmatic centres. This is consistent with the metal distribution patterns of nickel-copper-PGE sulphide mineralization associated with Nipissing gabbro and the inferred magmatic centre, in the Sudbury area (Innes and Colvine 1984). Crustal assimilation did take place along the roofs of individual sills and is marked in part by the presence of granophyre (Card and Pattison 1973; Lightfoot and Naldrett 1989). This assimilation along sites of emplacement is regionally developed, is not generally related to sites of sulphide concentration and therefore, is less likely to account for the regional metal distribution patterns.

A genetic relationship between Nipissing gabbro and hydrothermal veins is less likely. Vein types "1", "2", "3", "5" and "6" (*see* Hydrothermal Veins) formed in, and parallel to, 1.86 billion-year-old Penokean deformation structures (cf., Pearson 1980). These veins formed about 300 million years after the emplacement and cooling of the 2.2 billion-year-old Nipissing gabbro (cf., Pearson 1980; Andrews et al. 1986). Bornite-rich veins of type "4" constitute part of the copper-rich vein zone in the western part of the Huronian Supergroup (Innes and Colvine 1984).

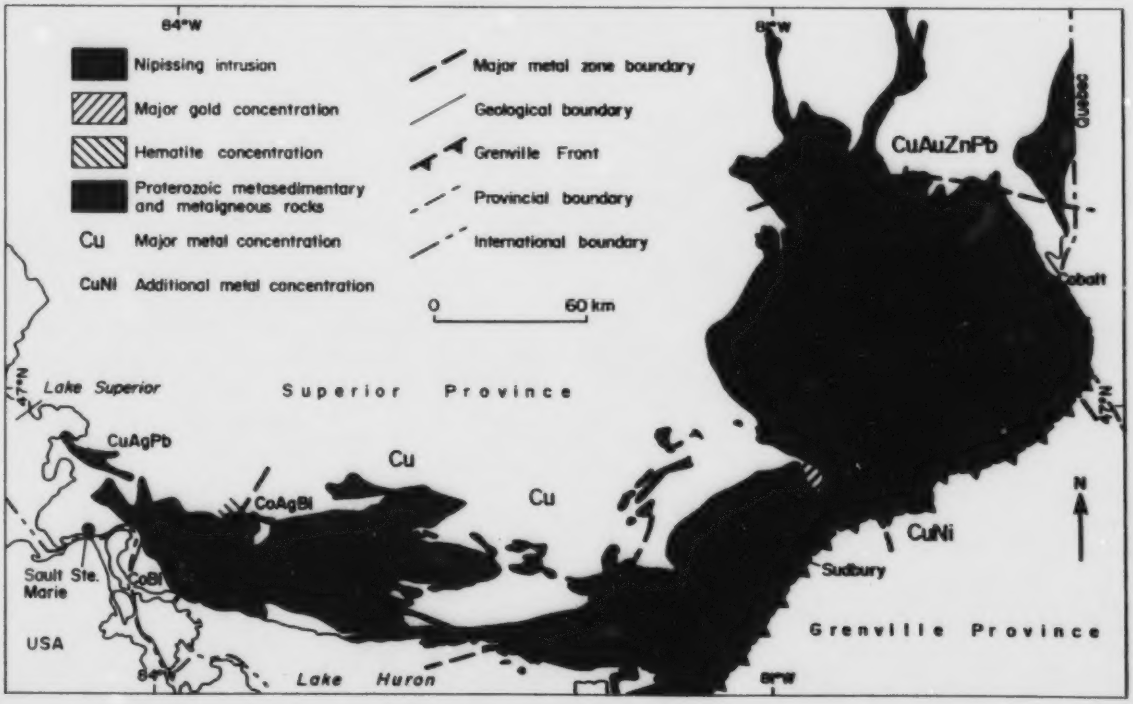

84°W
80°W
Nipissing intrusion
Major gold concentration
Hematite concentration
Proterozoic metasedimentary and metaigneous rocks
Cu Major metal concentration
CuNi Additional metal concentration
Major metal zone boundary
Geological boundary
Grenville Front
Provincial boundary
International boundary
0 60 km
CuAuZnPb
Quebec
Cobalt
47°N
47°N
Lake Superior
Superior Province
CuAgPb
Cu
Cu
CoAgBi
CoBi
Sault Ste. Marie
USA
N
CuNi
Sudbury
Grenville Province
84°W
Lake Huron
80°W

**Figure 23.17.** Distribution of metal assemblages associated with Nipissing gabbro in Proterozoic rocks in the area east of Lake Huron (*modified from* Innes and Colvine 1979, 1984).

These bornite-rich veins formed in 1.1 billion-year-old Keweenawan structures, at least 1.0 billion years after the emplacement of Nipissing gabbro. Gold- and copper-gold-bearing quartz-carbonate veins and attendant sodic metasomatic zones appear to have formed 500 million years after the emplacement of Nipissing gabbro during a 1.7 billion-year-old (Easton 1992) orogeny that affected principally the northwestern Grenville Province in Ontario. Hence, metal zoning involving the hydrothermal polymetallic veins in the Huronian Supergroup developed over as much as 1.1 billion years and cannot be simply attributed to Nipissing magmatism as implied by Innes and Colvine (1984). Nipissing gabbro appears to have played a mechanical role in the siting of the polymetallic vein deposits. A genetic relationship may exist between Nipissing gabbro and magmatic nickel-copper-PGE mineralization (cf., Andrews et al. 1986; Franklin et al. 1986; Lightfoot and Naldrett 1989) and the polymetallic skarn mineralization.

## Tectonic History and Relationship to Metallic Mineral Deposits

Excellent metallogenic reviews exist for Proterozoic rocks in the northern Great Lakes area (cf. Franklin 1970; Pearson 1980; Innes and Colvine 1984). However, additional critical spatial and temporal relationships between mineralization, faults, deformation fabrics, metamorphic and alteration mineral assemblages are required to describe a mineral deposit and to relate the formation of that deposit type to the tectonic history. The scarcity of these observations hinders the reconstruction of time and space relationships between the tectonic history and mineral deposits, and prohibits a rigorous testing of the depositional models (e.g., syngenetic versus epigenetic hydrothermal events). Regardless, general time and space relationships can be established between some of the deposit types and the tectonic history.

Paleoproterozoic and Mesoproterozoic rocks in the Lake Superior and Lake Huron areas evolved over a period of 1.5 billion years and experienced similar tectonic histories (Sutcliffe, this volume; Bennett et al., this volume). However, the tectonic history in the area east of Lake Huron is apparently more protracted than that in the Lake Superior area (Figure 23.18). Several distinct tectonic superepisodes, defined by the clustering of tectonic events in time, culminated in an orogeny. The tectonic superepisodes are separated by an apparent absence of tectonic activity (Figure 23.18). During this protracted tectonic history, a spectrum of mineral deposit types formed in diverse tectonic settings, during the tectonic superepisodes (Figure 23.18).

### 2.6 Ga: TERMINATION OF THE KENORAN OROGENY

One of the last accretionary events during the Neoarchean occurred at about 2.6 Ga in the western Lake Superior region, when a 2.7 billion-year-old granite-greenstone terrane (Superior Province) was juxtaposed against a 3.5 billion-year-old gneiss terrane (Minnesota River Valley gneiss terrane) along the Great Lakes Tectonic Zone (Peterman et al. 1980; Sims et al. 1980). The postulated eastward extension of this zone, in the Lake Huron region, is the Murray Fault (Sims et al. 1980; Zolnai et al. 1984). No metallic mineral deposits of significance appear to have formed in the northern Great Lakes region during this event, although, the large lode gold deposits in the southern Superior Province formed prior to or during this collisional event.

## Tectonic Summary—Lake Superior Region

Three distinct Proterozoic tectonic superepisodes, each temporally associated with the formation of metallic mineral deposits, are documented for the Lake Superior region (Sutcliffe, this volume): 1) deposition of Paleoproterozoic, 2.1 to 1.85 billion-year-old Animikie continental margin sedimentary prism, prior to, and during, the circa 1.86 billion-year-old Penokean Orogeny and emplacement of circa 1.9 billion-year-old carbonatite and alkaline complexes, in proximity to, and along, Proterozoic faults; 2) intrusion of circa 1.54 billion-year-old, Mesoproterozoic anorogenic granites and the deposition of the rift-related Sibley Group; and 3) Mesoproterozoic rifting along the Midcontinent Rift System between circa 1.11 to 1.09 billion years, during which the Keweenawan Supergroup was deposited.

## Tectonic Summary—Area East of Lake Huron

The Paleo- and Mesoproterozoic history in the area east of Lake Huron is more protracted than that documented for the Lake Superior region. At least 6 distinct tectonic superepisodes are documented: 1) rifting of the composite Archean craton at about 2.45 Ga and attendant erosion resulted in deposition of a northward-onlapping, shallow-water, clastic prism (Huronian Supergroup), upon a south-facing continental margin; emplacement of the 2.39 and 2.33 billion-year-old Murray and Creighton granitic plutons into lower Huronian Supergroup may have been related to the late stages of the rift event; 2) deformation of the Huronian Supergroup and the subsequent intrusion of 2.22 billion-year-old Nipissing gabbro; 3) the circa 1.86 billion-year-old Penokean Orogeny, during which the Huronian Supergroup clastic prism was overridden by an unknown, allochthonous terrane and underwent crustal thickening, metamorphism and thrust faulting. The emplacement of the 1.85 billion-year-old norite of the Sudbury Igneous Complex was coeval with the Penokean Orogeny. This magmatism was possibly triggered by a meteorite impact, an endogenic process, or a combination of both; 4) 1.74 billion-year-old anorogenic magmatism in the Killarney Magmatic Belt that may have been induced by the extensional phase of a circa 1.7 billion-year-old collisional orogeny that affected the northwestern part of the Grenville Province; 5) a period of 1.45 billion-year-old anorogenic magmatism within the Killarney Magmatic Belt that was coeval with 1.55 to 1.4 billion-year-old anorogenic magmatism in the Lake Nipigon and Wisconsin areas; 6) emplacement of the 1.24 billion-year-old, west-northwest-striking Sudbury diabase dike swarm that is deformed and metamorphosed within the Grenville Front Tectonic Zone, demonstrating that orogenic events, related to the Grenville Orogeny, had been initiated by 1.24 Ga in this part of the Grenville Orogen. The 1.2 to 1.0 billion-year-old Grenville Orogeny, evidence for which is recorded only a few kilometres north of the Grenville Front Tectonic Zone, involved northwest-directed crustal-scale thrusting, along southeast-dipping structures and extensive reworking of 1.8 to 1.35 billion-year-old crustal material.

## 2.5 TO 2.4 Ga: HURONIAN SUPERGROUP—A RIFT BASIN AND PASSIVE-MARGIN SUCCESSION

At about 2.45 Ga, the composite Archean craton was rifted along a series of east-striking structures. Some of these structures, such as the Flack Lake Fault, were newly created while others, such as the Great Lakes Tectonic Zone, were reactivated (Sims et al. 1980; Zolnai et al. 1984). Although correlations remain tenuous, incipient intracratonic rift development, in the eastern part of the area, may have triggered: 1) the emplacement of the 2.48 billion-year-old (Krogh et al. 1984) East Bull Lake, Shakespeare–Dunlop and related gabbro-anorthosite complexes; 2) eruption of the 2.45 billion-year-old Copper Cliff rhyolite, subaerial Thessalon Formation and slightly younger volcanic units to the east (Card and Hutchinson 1972; Innes 1977; Krogh et al. 1984); and 3) intrusion of the 2.45 billion-year-old (Heaman 1989) Matachewan dike swarm. Continued rifting and volcanism resulted in formation of a new continental margin along which sediments of the Huronian Supergroup were eventually deposited. The 2.45 billion-year-old (Heaman 1989) Matachewan dike swarm and the sedimentary rocks of the Huronian Supergroup, occupying the Cobalt Embayment, possibly represent the products of a rift-related failed arm (Fahrig 1987).

In the Huronian Supergroup, nickel-copper-PGE mineralization, spatially associated with the East Bull Lake and Shakespeare–Dunlop gabbro-anorthosite intrusions, developed during this period. Virtually all VMS-like sulphide deposits and occurrences outside of the Sudbury Basin formed in, or in close proximity to, the rift-related, submarine metavolcanic rocks of the Salmay Lake and Stobie formations. No VMS occurrences occur within the subaerial, metavolcanic rocks of the Thessalon Formation, although amygdule-filling chalcopyrite is common.

Deposition of a northward-onlapping, shallow-water, in part glaciogenic, conglomerate-pelite-arenite, clastic prism took place following the initial effusion of basaltic volcanism and the formation of the associated VMS deposits. This sedimentary prism is inferred to have developed upon a south-facing continental margin (Card et al. 1972; Zolnai et al. 1984). The important paleoplacer uranium deposits at Elliot Lake (Bennett et al., this volume), contained within quartz-pebble conglomerate units of the Matinenda Formation, accumulated early in this period. The Lorrain and lower Gordon Lake formations, part of which may have been deposited in a tropical environment (cf., Wood 1973; Chandler 1986), marked the end of this sedimentary period. The poorly documented shale-hosted copper mineralization in the Gowganda Formation, and the sediment-associated red-bed copper occurrences in the Lorrain Formation, formed during, or following, the accumulation of the basal Gordon Lake Formation. The red-bed-type copper mineralization, in the Lorrain and Gordon Lake formations, represents one of the oldest examples of this deposit type. The inferred tropical setting of the Gordon Lake Formation may have been a critical factor contributing to the localization of this copper mineralization (cf., Wood 1973; Chandler 1986).

## 2.4 TO 2.2 Ga: POST-HURONIAN PLUTONISM

Between 2.39 and 2.33 Ga (Frarey et al. 1982; Krogh et al. 1984), the Murray and Creighton granitic plutons were intruded into lower Huronian sediments near Sudbury. Following the emplacement of these granitic intrusions, but prior to the emplacement of the 2.22 billion-year-old (Corfu and Andrews 1986) Nipissing gabbro, partially lithified Huronian sediments were deformed (cf., Young 1983). The origin and tectonic significance of these structures and intrusions is enigmatic, and they may represent the transition between continental rifting and the Penokean Orogeny (Bennett et al., this volume).

Spatially related to the 2.22 billion-year-old (Corfu and Andrews 1986) Nipissing gabbro are magmatic, nickel-copper-PGE, sulphide deposits (Pearson 1980; Innes and Colvine 1984) and polymetallic contact metasomatic mineralization. The contact metasomatic mineralization developed where calcareous sediments of the Espanola Formation were contact metamorphosed by Nipissing gabbro. Silver-sulpharsenide veins in the Cobalt area, formed following the intrusion and cooling of the Nipissing gabbro possibly during the Penokean Orogeny.

## 2.2 TO 1.85 Ga: PENOKEAN OROGENY—AN ACCRETIONARY CONTINENTAL MARGIN

Between 1.95 and 1.80 Ga, in the Lake Superior area, collision along the Trans-Hudson Orogen resulted in the suturing of the Superior, Wyoming and Hearne Archean microcontinents (Hoffman 1989). No known metallic mineral deposits formed in Ontario during this event, although komatiite-associated nickel-copper-PGE and epigenetic gold mineralization formed elsewhere in the Trans-Hudson Orogen. Between 2.1 and 1.85 Ga, sedimentary rocks of the Animikie Basin formed as a continental margin sedimentary prism. At about 1.9 Ga, carbonatite and alkalic complexes were emplaced locally in Archean rocks, in proximity to, and along Proterozoic faults. Island arc volcanic rocks and associated VMS deposits of the allochthonous Wisconsin Magmatic Terrane developed between 1.9 and 1.75 Ga.

The Penokean Orogeny, in the western and southern Lake Superior area, resulted from accretion of the Wisconsin island-arc terrane onto the Archean craton and associated miogeoclinal sediments (Marquette Range Supergroup—Animikie Group), along a southward-dipping

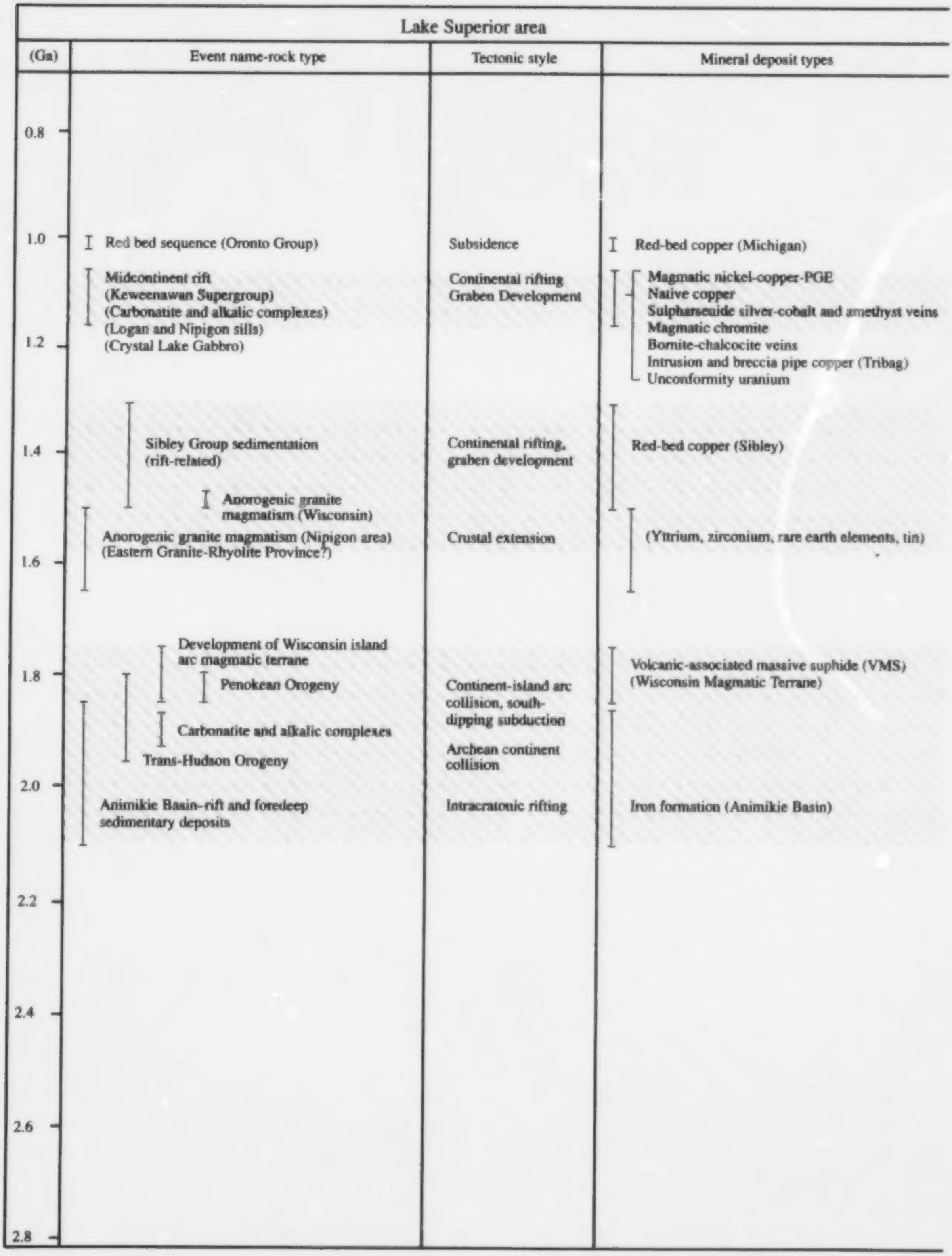


**Figure 23.18.** Synthesis of tectonic events that affected Proterozoic rocks in the northern Great Lakes area and the types of coeval metallic mineral deposits that occur in Ontario. Shaded areas indicate periods of more intense or discrete orogenic activity.

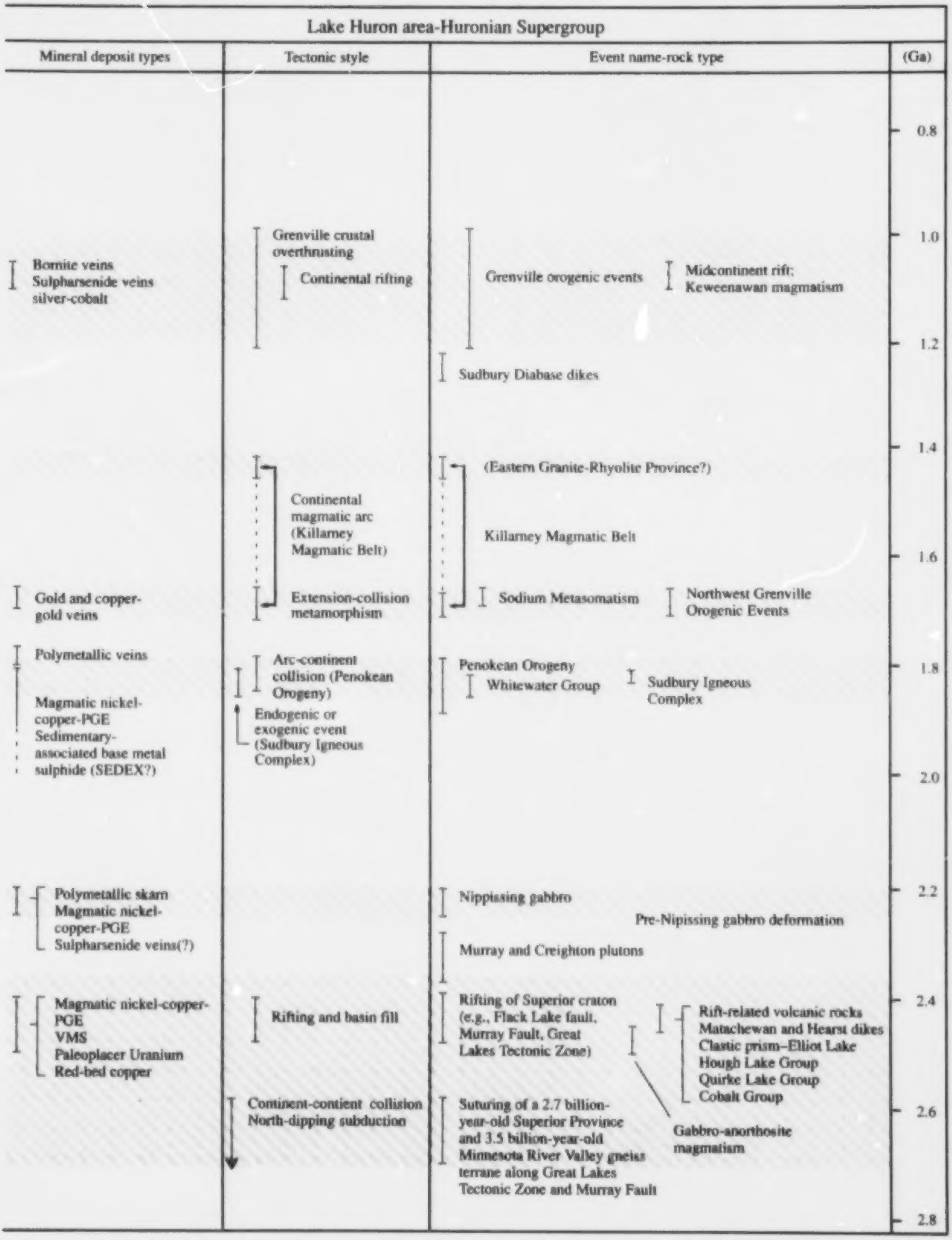


**Figure 23.18.** Synthesis of tectonic events that affected Proterozoic rocks in the northern Great Lakes area and the types of coeval metallic mineral deposits that occur in Ontario. Shaded areas indicate periods of more intense or discrete orogenic activity.

subduction zone (Cambray 1978; Larue 1983; Zolnai et al. 1984). This orogenic collisional event occurred between 1.9 and 1.76 Ga (Southwick et al. 1988), with most intense activity having occurred between 1.87 and 1.85 Ga (Van Schmus 1976; Van Schmus and Bickford 1981; Bickford et al. 1986). The 1.86 to 1.835 billion-year-old Niagara Fault Zone is interpreted as the present-day expression of the collision boundary between the 2 terranes (Cambray 1978; Larue 1983; Barovich et al. 1989). In the Huronian Supergroup, the orogeny involved northward overriding of the Huronian Supergroup clastic prism by an unknown, allochthonous terrane along north-verging thrust faults and reverse movements along reactivated normal faults, such as the Flack Lake and Murray faults (Dickin and McNutt 1989). The Huronian continental margin, south of the Murray Fault, was depressed to midcrustal levels, metamorphosed and was subsequently thrust northward and juxtaposed against lower metamorphic grade rocks (Zolnai et al. 1984).

At about 1.85 Ga (Krogh et al. 1982), the norites of the Sudbury Igneous Complex were emplaced. Associated with the igneous Sublayer rocks are the nickel-copper-PGE sulphide deposits of the Sublayer unit. Granophyric phases of the norite intrude the lower part of the Onaping Formation; hence, deposition of the basal part of the Onaping Formation, preserved only within the Sudbury Basin, postdates the 2.22 billion-year-old (Corfu and Andrews 1986) Nipissing gabbro and predates the 1.85 billion-year-old (Krogh et al. 1982) norites of the Sudbury Igneous Complex. Textures consistent with shock metamorphism of zircons in the Onaping Formation are inferred to have formed at about 1.836±14 Ga (Krogh et al. 1984). These shock metamorphic textures are attributed to an impact origin (Dietz 1964; Dence 1972; *see* Dressler, this volume), although some debate continues (cf., Muir 1984). Alternatively, all or parts of the Onwatin and Chelmsford formations, preserved as the Whitewater Group in the Sudbury Basin, may represent synorogenic foredeep sediments (Young 1983) comparable in age and tectonic setting with the Animikie sediments (*see* Figure 23.18). The sedimentary-associated (SEDEX), massive, base metal sulphide deposits, at the base of the carbonate-rich Vermilion Member, formed during this period.

In the Huronian Supergroup, east-striking, quartz-carbonate-chalcopyrite-pyrite, gold-chalcopyrite, copper-cobalt-(bismuth)-(uranium)-(nickel) and minor zinc-lead-silver-copper veins occur spatially related to Nipissing gabbro, commonly cutting rocks of the Gowganda Formation. The veins occur within, and strike parallel to, major Penokean structures such as the hinge zone of the Quirke Syncline. This indicates that the veins formed in fractures during, or after, the Penokean Orogeny (Pearson 1980), at least 300 million years after the intrusion of Nipissing gabbro. This temporal relationship precludes a genetic relationship between Nipissing gabbro and the hydrothermal veins.

## 1.75 TO 1.30 Ga: KILLARNEY MAGMATIC BELT, ANOROGENIC MAGMATISM AND RIFT-RELATED SEDIMENTATION

The wedge-shaped Killarney Granitic Complex contains 1.74 and 1.45 billion-year-old granitoid rocks that intruded the Huronian Supergroup previously deformed during the Penokean Orogeny (Davidson 1986). Preliminary investigations indicate that at least the Killarney granite pluton is an anorogenic pluton (*see* Easton, this volume). The complex records intrusive, metamorphic and structural events temporally between the Penokean and Grenville orogenies (Davidson 1986). The 1.74 billion-year-old anorogenic magmatism in the Killarney Magmatic Belt may reflect the extensional phase of a circa 1.7 billion-year-old collisional orogeny that affected the northwestern part of the Grenville Province (Easton 1992). No metallic mineralization is known to be spatially related to these anorogenic intrusions. However, the spatial and temporal association between gold and copper-gold hydrothermal veins, attendant circa 1.7 billion-year-old (Schandl et al. 1991) sodium metasomatic alteration of Huronian metasedimentary rocks (Meyer et al. 1988, 1989, 1990; Gates 1991) and a proposed 1.7 billion-year-old continental collision orogeny that affected the northwestern part of the Grenville Province in Ontario (Easton 1992) indicates that all may have developed during the same, Post-Penokean orogenic event.

A period of 1.45 billion-year-old anorogenic magmatism within the Killarney Magmatic Belt is coeval with 1.55 to 1.4 billion-year-old anorogenic magmatism in the Lake Nipigon and Wisconsin areas. These intrusions, possibly related to the Eastern Granite-Rhyolite Province, may indicate pre-Grenville, crustal extension (*see* Sutcliffe, this volume). Although the intrusions contain anomalous amounts of yttrium, zirconium, rare earth elements and tin, no deposits are known. Accumulation of the Sibley Group sedimentary rocks took place between 1.5 and 1.3 Ga, possibly related to rifting along the Black Sturgeon fault. Unconformity- related uranium mineralization and red-bed copper mineralization in the Sibley Group developed during this period.

## 1.30 TO 1.0 Ga: GRENVILLE OROGEN AND MIDCONTINENT RIFT

At 1.24 Ga, the west-northwest-striking Sudbury dike swarm was emplaced into the Huronian Supergroup. Bethune and Davidson (1988) mapped deformed and metamorphosed Sudbury diabase dikes into the Grenville Front Tectonic Zone, demonstrating that orogenic events, possibly related to the Grenville Orogeny, had been initiated by 1.24 Ga in this part of the Grenville Orogen (Easton, this volume).

In the Lake Superior area, at approximately 1.10 Ga, the crust experienced a series of extensional and magmatic events including the development of the Mesoproterozoic

Midcontinent Rift System (Behrendt et al. 1988 and references therein). Northwest-striking bornite-chalcocite-native copper and zinc-lead-silver-copper veins in the Sault Ste. Marie-Batchawana area, and galena-sphalerite-chalcopyrite-pyrite-barite, silver sulpharsenide and amethyst veins in the Lake Superior area, formed in and adjacent to Keweenawan fault zones. These vein sets are younger than the other vein sets localized along, or near, Penokean structures in the Huronian Supergroup (Pearson 1980). The copper-rich, sulphur-poor mineral assemblages of the bornite-bearing veins are consistent with interaction between hydrothermal fluids and the associated sulphur-poor Keweenawan basalts (Pearson 1980). The basalts may have acted as a copper source (cf., Heslop 1970).

Several other metallic mineral deposit types formed in the Lake Superior area during and following the Keweenawan magmatism. Native copper was precipitated in amygdules and chalcocite-quartz-calcite veins in Keweenawan basalt flows. Red-bed (shale-hosted) copper mineralization formed in the Nonesuch Shale and Freda Formation. Magmatic nickel-copper-PGE sulphide mineralization formed in Keweenawan gabbro intrusions such as the Duluth Gabbro and the Crystal Lake Complex. Magmatic nickel-copper-palladium mineralization formed locally in the Port Coldwell alkalic complex. In the Lake Huron area, the Tribag breccia- and Jogran felsic intrusion-associated copper mineralization formed during Keweenawan tectonic events.

During the 1.2 to 1.0 billion-year-old Grenville Orogeny, northwest-directed, crustal-scale thrusting along southeast-dipping structures, over a region of the Huronian Supergroup and extensive reworking of 1.5 to 1.3 billion-year-old crustal material reflects a major collisional event (cf., Green et al. 1988). No significant mineralization in the Huronian Supergroup appears to have developed during these tectonic events.

## Summary

During the complex and protracted tectonic history of the Proterozoic rocks in the northern Great Lakes area of Ontario, several metallic mineral deposit types developed episodically in response to different tectonic events (*see* Figure 23.18). Magmatic nickel-copper-PGE mineralization formed: 1) at 2.48 Ga associated with gabbro-anorthosite intrusions; 2) at 2.2 Ga associated with Nipissing gabbro; 3) at 1.85 Ga associated with the Sudbury Igneous Complex; and 4) at 1.0 Ga associated with Keweenawan gabbro and alkalic complexes. Red-bed copper mineralization formed: 1) in the *circa* 2.5 to 2.4 billion-year-old lower Huronian Supergroup; 2) in the 1.5 to 1.3 billion-year-old Sibley Group sedimentary rocks; and 3) in the 1.1 billion-year-old sedimentary rocks in the upper Keweenawan succession. Volcanic-associated, massive, base metal sulphide deposits formed at *circa* 2.45 Ga in mafic metavolcanic rocks in the lower part of the Huronian Supergroup and again at *circa* 1.85 to 1.75 Ga in dacite rhyolite successions of the Wisconsin Magmatic Terrane. Sedimentary-associated, massive base metal sulphide deposits formed in the Whitewater Group at approximately 1.85 Ga. These sedimentary-associated deposits share some characteristics of SEDEX deposits. Hydrothermal veins, formed in structures, formed during the *circa* 1.85 billion-year-old Penokean Orogeny, a 1.7 billion-year-old orogeny that affected the northwest part of the Grenville in Ontario and in 1.1 billion-year-old Keweenawan structures. This episodic formation of similar deposit types reflects the protracted sedimentary, magmatic, metamorphic and deformation history of the northern Great Lakes region over approximately 1.5 billion years.

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## Chapter 24

# Metallogeny of the Grenville Province

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## Abstract

This chapter reviews the tectonic setting of the major mineral deposits of the Grenville Province. Many of Ontario's first mines were located in the Central Metasedimentary Belt of the Grenville Province, and significant production of iron, zinc, molybdenum, uranium and lead occurred, primarily at the turn of the century. Copper, nickel and gold prospects are also common in the Central Metasedimentary Belt, and limited gold production has occurred. Industrial mineral production has been historically important and continues as a mainstay of Grenville Province mineral production primarily within the Central Metasedimentary Belt, and includes graphite, nepheline syenite, talc, tremolite, calcite and dolomite marbles, roofing granules, flagstone and building stone.

The first comprehensive synthesis of the mineral potential of the Central Gneiss Belt is offered here. Paragneiss-rich domains within the Algonquin Terrane host graphite deposits and potentially host SEDEX base metal deposits. The Parry Sound Domain, with its abundance of mafic metaigneous rocks and paragneisses, hosts numerous gold and copper showings. Industrial mineral potential of the Central Gneiss Belt includes flagstone and building stone resources, graphite and the use of anorthosite as an alumina source.

The mineral potential of the Central Metasedimentary Belt is significantly higher than the Central Gneiss Belt, although some commodities are concentrated in specific terranes and domains, a reflection of the stratigraphy and tectonic diversity of the region. Mineralization within the Central Metasedimentary Belt appears to have occurred mainly in extensional regimes during 4 periods: 1) copper-nickel-platinum group element (PGE) mineralization in ca. 1290 to 1270 million-year-old mafic intrusions, probably formed during oceanic crust generation in a marginal basin environment; 2) iron-titanium, iron-skarn, iron-sulphide and iron formation, stratiform carbonate-hosted zinc, and minor base metal mineralization related to volcanism, sedimentation and emplacement of Lavant Suite and Methuen Suite gabbroic and granitic plutons in the ca. 1255 to 1240 Ma interval, again, probably in a marginal basin. Many important industrial mineral deposits developed at this time in the contact aureoles of Lavant Suite and Methuen Suite plutons; 3) gold mineralization, possibly related to local extensional environments that developed post–Elzevirian Orogeny at ca. 1180 to 1155 Ma. Graphite deposits in the Central Metasedimentary Belt may also have formed at this time; and 4) uranium, rare element and molybdenum deposits related to granite pegmatite injection and regional metasomatism in the Bancroft Terrane post–Ottawan Orogeny at ca. 1060 to 1030 Ma. In addition to these 4 main mineralizing periods, a variety of industrial mineral deposits in the Mazinaw Terrane formed during Ottawan Orogeny regional metamorphism at ca. 1150 to 1090 Ma.

To date, most mineral deposits, including past- and present-producers found within the Grenville Province, are small in comparison to other regions (e.g., Superior Province). The authors know of no geologic reason for the absence of large ore bodies from the Grenville Province, and it is the authors' view that significant deposits remain to be discovered.

## INTRODUCTION

The Grenville Province has had a long history of mineral exploration and development, extending from the 1860s to the present. The first discovery of gold in Ontario was made in 1866 near Eldorado (Figure 24.1), and subsequently, iron, copper, lead, zinc, molybdenum and uranium were produced from Grenville Province mines (Figures 24.2 and 24.3). In addition, the Grenville Province hosts a variety of industrial mineral deposits which also have a long history of exploitation, notably graphite, nepheline syenite, talc, calcite and dolomite (Figures 24.4 and 24.5). Most metallic and industrial mineral deposits in the Grenville Province occur in the Central Metasedimentary Belt in southeastern Ontario and neighbouring parts of Quebec and New York State (Figure 24.6). With the exception of the Parry Sound area, few metallic mineral deposits occur in the Central Gneiss Belt (Carter 1984). In this chapter, the geologic controls on the distribution of mineral deposits within the Grenville Province, and the potential for further mineral exploration and development will be explored.

## GEOLOGICAL FRAMEWORK

The geological and tectonic framework of the Grenville Province has been outlined in the chapter entitled "The Grenville Province and the Proterozoic History of Central and Southern Ontario" (*see* Easton, this volume), and familiarity with that chapter is assumed. Briefly, the Grenville Province is subdivided into the Central Gneiss Belt and the Central Metasedimentary Belt (*see* Figure 24.6). The Central Gneiss Belt consists mainly of Archean to Mesoproterozoic quartzofeldspathic gneisses, with subordinate paragneiss, which have been metamorphosed to upper amphibolite and locally, granulite facies. The Central Gneiss Belt has been subdivided into a number of lithotectonic terranes and domains, as outlined by Easton

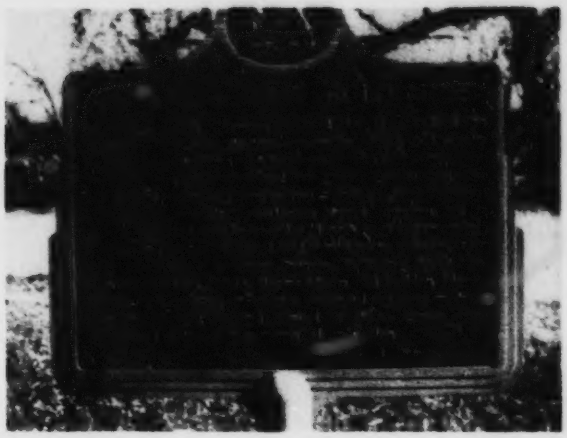


**Figure 24.1.** Historic plaque marking Ontario's first gold mine, discovered in 1866 at Eldorado, Ontario.

**Figure 24.2.** Ore dump at the Radenhurst–Caldwell magnetite deposit, Mazinaw Terrane, Central Metasedimentary Belt. Many small mines that operated at the turn of the century in the Grenville Province are almost completely overgrown, as shown in this example.

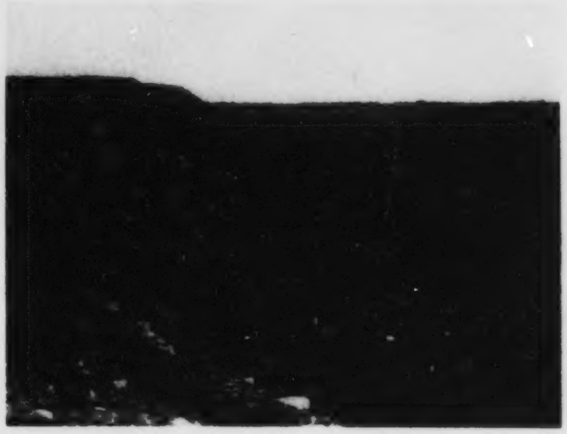


**Figure 24.3.** View of the Marmoraton open pit mine contact metasomatic magnetite deposit at Marmora, Ontario. The Marmoraton Mine deposit was discovered during an aeromagnetic survey of the area, and operated from 1955 to 1978.

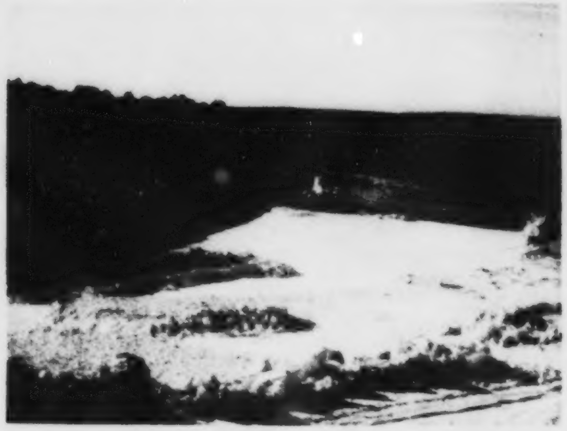


**Figure 24.4.** View of the calcite marble quarry near Tatlock, Ontario. The marble is used primarily for producing very fine-grained calcium carbonate products. Processing is done in Perth, 40 km distant by road.

(this volume) and in Figure 24.6. The part of the Central Gneiss Belt that lies within 20 to 30 km of the Grenville Front is termed the Grenville Front Tectonic Zone and consists of deformed and metamorphosed rocks derived from the Southern and Superior provinces and the Killarney Magmatic Belt (*see* Figure 24.6). The Central Gneiss Belt is separated from the Central Metasedimentary Belt by a 2 to 20 km wide zone of ductile mylonites, the Central Metasedimentary Belt Boundary Zone (*see* Figure 24.6).

The Central Metasedimentary Belt consists of Mesoproterozoic marbles, metavolcanic rocks and clastic metasedimentary rocks intruded by several compositionally diverse suites of pre-, syn- and posttectonic plutons (*see* Table 19.16, this volume). Metamorphic grade in the area ranges from greenschist to granulite facies (*see* Figure 19.15, this volume). The Central Metasedimentary Belt is also subdivided into a series of distinct lithotectonic terranes and domains (*see* Figure 24.6). The Central Metasedimentary Belt has been affected by at least 2 orogenic events (*see* Easton, this volume for complete discussion); the Elzevirian Orogeny (ca. 1230 to 1180 Ma) and the Ottawan Orogeny (ca. 1160 to 1070 Ma). The Ottawan Orogeny has in the past generally been equated with the Grenville Orogeny, although the Elzevirian Orogeny may be largely responsible for establishing the tectonic and metamorphic framework of the Central Metasedimentary Belt.

## DEPOSIT CLASSIFICATION

Carter (1984), Carter and Colvine (1985), Sangster (1970) and Sangster and Bourne (1982) outlined a metallic mineral deposit classification for the Central Metasedimentary Belt. Easton et al. (1986) extended this classification to industrial mineral deposits. These classifications served as the basis

**Figure 24.5.** Headframe of the Canada Talc Mine at Madoc, Ontario. Dolomite is also quarried at this site, as shown in the foreground. The Henderson Shaft shown here has been in operation since 1893.

for the classification used in this volume (Table 24.1). Sulphide mineral assemblages are listed using the standard Ontario Geological Survey abbreviations.

The concept of stratigraphic and lithologic control on mineralization, noted by Carter (1984) and Carter and Colvine (1985), is further emphasized herein, because the major differences in stratigraphy, plutonism and tectonism present between the different Central Metasedimentary Belt and Central Gneiss Belt terranes and domains are reflected in the distribution of mineral deposits (Tables 24.2 and 24.3). Specific types of mineral deposits are absent from some domains in both the Central Metasedimentary and Central Gneiss belts because of the lack of critical lithologies or depositional and tectonic events which played an important role in the mineralization process. In addition, for the first time, the metallogeny of the Central Gneiss Belt is evaluated; a reflection of recent advances in our knowledge of this region over the past decade.

### Importance of Known Metallic Deposits

Few major metallic mineral deposits of any type have been discovered in the Grenville Province of Ontario, with the possible exception of the Marmoraton iron and the Long Lake zinc mines (Carter 1984). Occurrences and deposits are small in comparison to similar deposits present in parts of the Superior Province (*see* Fyon, Breaks et al., this volume). The apparent scarcity of large metallic mineral deposits poses several important questions. Does the absence of major metallic mineral deposits reflect aspects of the tectonic history of the Grenville Province or has the incomplete geological data base and complex land tenure situation within the Grenville Province of Ontario discouraged exploration? Has the lack of comprehensive mineral deposit descriptions and an understanding of ore-forming mechanisms, in part a reflection of the incomplete geological data base, hindered the formulation of exploration models applicable to the Grenville Province? Or have the high grades of metamorphism and burial to deep crustal levels during the Grenville Orogeny, particularly in the Central Gneiss Belt, flushed metals to higher crustal levels?

Analysis of the tectonic history of the Grenville Province of Ontario (*see* Easton, this volume) indicates that igneous and sedimentary rocks in the Grenville Province formed in disparate tectonic settings. Some of the interpreted tectonic settings represented in the Grenville Province of Ontario are known to contain metallic mineral deposits in older (e.g., Superior Province; *see* Fyon, Breaks et al., this volume) and modern environments. Therefore, it is contended herein that the absence of major metallic mineral deposit types in the Grenville Province cannot be attributed simply to aspects of the tectonic and metamorphic history of the Grenville Province, as presently understood. Consequently, major metallic mineral deposits may potentially occur throughout the Grenville Province, particularly within the Central Metasedimentary Belt.

## MINERAL DEPOSITS IN THE CENTRAL GNEISS BELT

Little modern mineral exploration has been conducted within the Central Gneiss Belt and the Grenville Front Tectonic Zone, in part because of the perception that high-grade gneiss terranes have little mineral potential. Lack of detailed geologic maps and mineral deposit models have also discouraged mineral exploration. Nevertheless, our understanding of the tectonic architecture of the region allows establishment of a broad framework for mineral exploration within the Central Gneiss Belt.

Flagstone is the major commodity that is not domain-specific, as the best flagstone prospects are found in transposed and straight gneiss zones associated with terrane or domain boundaries. Also associated with terrane or domain boundaries are ca. 1170 million-year-old mafic and ultramafic bodies (metabasites and pseudoeclogites, cf., Davidson 1991) which have recently been explored for PGE, copper and gold mineralization (Garland and Villard 1990).

Given the lack of detailed knowledge of much of the Central Gneiss Belt, it is more useful to discuss the metallogenic characteristics and potential of each terrane in a broad context. Table 24.2 outlines the mineral deposits and potential of the various terranes of the Central Gneiss Belt. Additional details on specific deposits within terranes are found elsewhere in this volume (*see* Easton, this volume).

### Grenville Front Tectonic Zone

The Grenville Front Tectonic Zone is divided into 3 segments reflecting the geology of the shield west and north of the Grenville Front. One characteristic of much of the region north of the French River, both in the Grenville Front Tectonic Zone and in the Nepewassi Domain in particular (*see* Figure 24.6), is the lack of closely spaced jointing in many of the metaplutonic rocks. This jointing pattern is

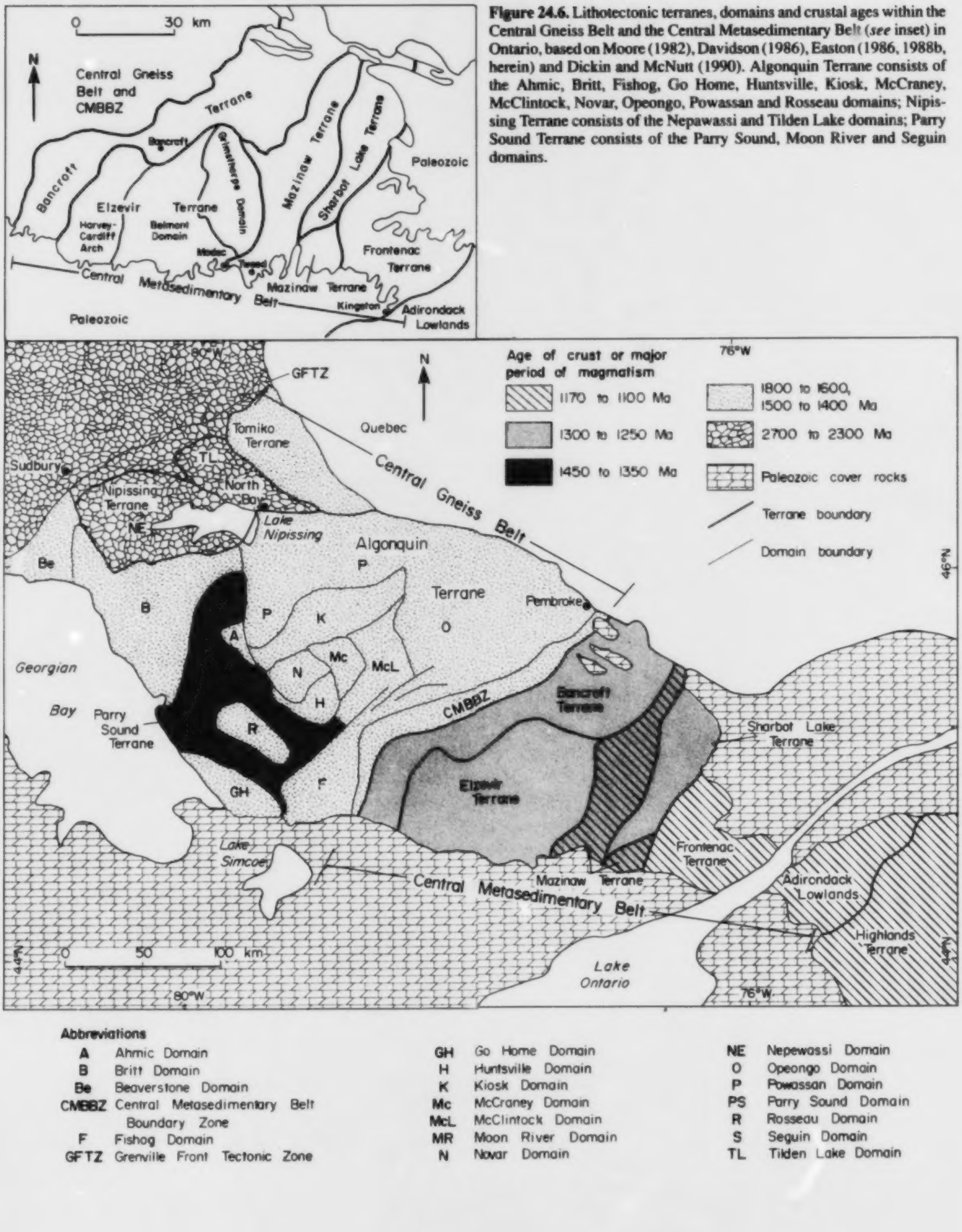

0 30 km
N
Central Gneiss Belt and CMBBZ
Terrane
Bancroft
Bancroft
Grimsthorpe Domain
Mazinaw Terrane
Sharbot Lake Terrane
Paleozoic
Elzevir Terrane
Harvey-Cardiff Arch
Belmont Domain
Madoc
Tweed
Frontenac Terrane
Central Metasedimentary Belt
Mazinaw Terrane
Kingston
Adirondack Lowlands
Paleozoic
80°W
GFTZ
N
Age of crust or major period of magmatism
1170 to 1100 Ma
1300 to 1250 Ma
1450 to 1350 Ma
1800 to 1600, 1500 to 1400 Ma
2700 to 2300 Ma
Paleozoic cover rocks
Terrane boundary
Domain boundary
76°W
Tomiko Terrane
Quebec
Sudbury
TL
North Bay
Nipissing Terrane
NE
Lake Nipissing
Central Gneiss Belt
Algonquin
P
Terrane
Be
B
P
K
O
Pembroke
46°N
A
Georgian
Bay
Parry Sound Terrane
N
Mc
McL
H
R
CMBBZ
Bancroft Terrane
Sharbot Lake Terrane
Elzevir Terrane
GH
F
Lake Simcoe
Frontenac Terrane
Mazinaw Terrane
Adirondack Lowlands
Central Metasedimentary Belt
Highlands Terrane
44°N
0 50 100 km
Lake Ontario
80°W
76°W
44°N

**Figure 24.6.** Lithotectonic terranes, domains and crustal ages within the Central Gneiss Belt and the Central Metasedimentary Belt (*see* inset) in Ontario, based on Moore (1982), Davidson (1986), Easton (1986, 1988b, herein) and Dickin and McNutt (1990). Algonquin Terrane consists of the Ahmic, Britt, Fishog, Go Home, Huntsville, Kiosk, McCraney, McClintock, Novar, Opeongo, Powassan and Rosseau domains; Nipissing Terrane consists of the Nepawassi and Tilden Lake domains; Parry Sound Terrane consists of the Parry Sound, Moon River and Seguin domains.

**Abbreviations**

| | |
|---|---|
| A | Ahmic Domain |
| B | Britt Domain |
| Be | Beaverstone Domain |
| CMBBZ | Central Metasedimentary Belt Boundary Zone |
| F | Fishog Domain |
| GFTZ | Grenville Front Tectonic Zone |
| GH | Go Home Domain |
| H | Huntsville Domain |
| K | Kiosk Domain |
| Mc | McCraney Domain |
| McL | McClintock Domain |
| MR | Moon River Domain |
| N | Novar Domain |
| NE | Nepewassi Domain |
| O | Opeongo Domain |
| P | Powassan Domain |
| PS | Parry Sound Domain |
| R | Rosseau Domain |
| S | Seguin Domain |
| TL | Tilden Lake Domain |

**Table 24.1.** Classification of metallic and industrial mineral deposits in the Central Metasedimentary Belt (*modified from* Carter 1984; Easton et al. 1986).

**Metallic Mineral Deposits**

Volcanic- and Sediment-Associated Deposits

1. strata-bound, volcanic- and sediment-hosted pyritic and base metal-bearing sulphide deposits
2. strata-bound, carbonate-hosted zinc deposits (possibly several different genetic types are present)
3. iron formation

Magmatic Deposits (several plutonic suites represented)

4. copper-nickel sulphide deposits
5. iron-titanium oxide deposits

Epigenetic Deposits

6. contact metasomatic iron deposits
7. gold-quartz ± carbonate veins
8. metasomatic uranium, thorium, molybdenum skarn and vein deposits
9. barium ± fluorite ± strontium ± lead veins

Supergene Deposits

10. hematite developed by weathering of type 2b and 3a deposits

**Industrial Mineral Deposits**

1. Original Sedimentary and Volcanic Deposits
   - a. calcium carbonate
   - b. dolomite
   - c. barite
   - d. anhydrite
   - e. basalt
   - f. graphite
2. Metasomatic Deposits
   - a. quartz (in veins)
   - b. talc-tremolite
   - c. hematite
   - d. mica
   - e. wollastonite
3. Metamorphic Deposits
   (all have some degree of primary compositional control)
   - a. talc-tremolite
   - b. marble
   - c. graphite
   - d. garnet
   - e. zircon
   - f. beryl
   - g. apatite
   - h. brucite
   - i. mica
   - j. sillimanite
   - k. wollastonite
   - l. stone
   - m. corundum
4. Igneous Deposits
   - a. feldspar
   - b. quartz
   - c. nepheline
   - d. zircon
   - e. beryl
   - f. mica
   - g. building stone
   - h. apatite
5. Supergene Deposits
   - a. vermiculite
   - b. hematite
6. Epigenetic (Post–Grenville) Vein Deposits
   - a. fluorite
   - b. barite
   - c. celestite

more favourable than that observed south of the French River, and accounts for the numerous building stone prospects in the Grenville Province south of Sudbury (Lacey 1989) compared to the western Central Gneiss Belt (Marmont 1991).

## SEGMENT 1

Segment 1 consists of reworked Killarney Magmatic Belt granitic rocks. In the Killarney Magmatic Belt proper, these high-level granitic complexes intrude coeval felsic metavolcanic rocks (van Breemen and Davidson 1988a). On the Grenville Province side, however, deeper levels of these intrusions are exposed (Davidson and Bethune 1988). Supracrustal rocks in the area consist mainly of metamorphosed quartz arenites and quartzose paragneisses. Metallic mineral potential of this area is low. Lacey (1989) summarized the building stone potential of the area. Killarney Magmatic Belt granitoids on both sides of the Grenville Front show potential as building stone. Near Sudbury, kyanite-bearing schists in Dill and Dryden townships may have commercial potential (Pearson 1962).

## SEGMENT 2

Segment 2 consists of reworked Southern Province rocks, and thus, has the same potential as adjacent Southern Province rocks as described by Fyon, Bennett and Jackson (this volume). Southeast of Lake Wanapitei, the Grenville Front truncates a belt of gold-bearing quartz-arsenopyrite-pyrite veins with associated sodium metasomatism (*see* Figure 23.7, this volume). Deformed equivalents of these veins can be expected to occur southeast of the Grenville Front.

Near River Valley, metagabbroic and amphibolite bodies are common, and, in part, represent metamorphosed equivalents of the ca. 2490 million-year-old (Krogh et al. 1984) East Bull Lake-type gabbro-anorthosite intrusions of the Southern Province known to host nickel-copper-platinum group element mineralization (Peck and James 1989; *see* Fyon, Bennett and Jackson, this volume). Favourable zones in the East Bull Lake and Shakespeare–Dunlop gabbro-anorthosite intrusions include mixed nodular anorthosite-gabbro zones marginal to the intrusions (Peck and James 1989). Similar mineralization is possible in the River Valley and Red Deer gabbroic anorthosites south of the Grenville Front.

The River Valley Anorthosite has been quarried for building stone for many years, producing an attractive stone called Dana Black. This quarry is hosted in little deformed or metamorphosed anorthosite 1 km south of the Grenville Front. Lacey (1989) described the history of this property and several prospects located in more intensely deformed and metamorphosed parts of the River Valley Anorthosite. The deformed stones are black and white in colour, and are similar in appearance to Parry Sound area anorthosites. The anorthosites may also serve as potential alumina sources (Dolan et al. 1991).

## SEGMENT 3

Segment 3 consists of reworked Superior Province rocks, mainly the southern part of the Pontiac Subprovince (*see*

**Table 24.2.** Summary chart showing mineral potential of Central Gneiss Belt terranes and domains.

| Terrane/Domain | Age Range (Ma) | Dominant Rock Types | Mineral Deposit Types Present and Potential |
|---|---|---|---|
| **Grenville Front Tectonic Zone** | | | |
| Beaverstone Domain | Mesoproterozoic | granitic orthogneisses, metamorphosed quartz arenite and paragneiss | building stone, sillimanite |
| Segment 2 | Paleoproterozoic<br>Archean? | migmatitic gneisses, mafic orthogneisses | anorthosite as building stone and alumina source, kyanite, gold-arsenopyrite mineralization associated with metasomatism in Southern Province, nickel-copper-PGE associated with meta-Nipissing diabase dikes and metagabbroic and meta-anorthositic intrusions |
| Segment 3 | Archean | migmatitic gneisses, granular gneisses, minor orthogneiss | unknown? |
| **Nipissing Terrane** | (ca. 2700–1400) | | |
| Nepewassi Domain | Archean<br>Paleoproterozoic<br>Mesoproterozoic | migmatitic gneisses<br>granitic ortho- and paragneisses<br>tonalitic, monzonitic, granitic and anorthositic orthogneisses | low potential<br>SEDEX deposits?, paleoplacer deposits<br>building stone, meta-anorthosite as alumina source |
| Tilden Lake Domain | chiefly Archean? | migmatitic gneisses, paragneiss, minor orthogneiss | unknown? |
| **Tomiko Terrane** | (ca. 2500–1250)<br>Mesoproterozoic, possibly deposited on Archean basement | metamorphosed quartz arenite, paragneiss, minor iron formation, granitic orthogneisses | iron formation, SEDEX deposits, vermiculite, stone, mica, brucite |
| **Algonquin Terrane** | (ca. 1800–1450) | | |
| Ahmic Domain | Mesoproterozoic | dominantly orthogneiss | unknown |
| Britt Domain | Mesoproterozoic | paragneiss and ca. 1680 and 1450 Ma orthogneisses | building stone, flagstone in shear zones, SEDEX deposits in paragneisses? |
| Fishog Domain | Mesoproterozoic | dominantly orthogneiss | minor gold-copper mineralization in oldest orthogneisses, building stone, flagstone in internal shear zones |
| Go Home Domain | Mesoproterozoic | para- and orthogneiss | SEDEX deposits in paragneisses?, building stone and flagstone |
| Huntsville Domain | Mesoproterozoic | ortho- and paragneiss | flagstone, building stone |
| Kiosk Domain | Mesoproterozoic | para- and orthogneiss | graphite, SEDEX deposits? |
| McCraney Domain | Mesoproterozoic | orthogneiss | unknown |
| McClintock Domain | Mesoproterozoic | orthogneiss | unknown |
| Novar Domain | Mesoproterozoic | orthogneiss | unknown |
| Opeongo Domain | Mesoproterozoic | orthogneiss | unknown |
| Powassan Domain | Mesoproterozoic | ortho- and paragneiss | building stone, graphite, SEDEX deposits |
| Rosseau Domain | Mesoproterozoic | orthogneiss | building stone |
| **Parry Sound Terrane** | (ca. 1450–1350) | | |
| Moon River Domain | Mesoproterozoic | migmatitic gneisses | unknown |
| Parry Sound Domain | Mesoproterozoic | mafic gneisses, orthogneisses, paragneiss, metamorphosed quartz arenite and marble | gold-copper-PGE in mafic metaigneous rocks, meta-anorthosite as alumina source, flagstone in Parry Sound shear zone, marble as agricultural lime source |
| Seguin Domain | Mesoproterozoic | migmatitic gneisses | unknown |

**Table 24.3.** Summary chart showing mineralization spatially or genetically related to Central Metasedimentary Belt plutonic suites as defined in Chapter 19. Table 19.16 (*see* Easton, this volume) lists major rock types found in each suite.

| Rock Suite | Age Range (Ma) | Mineralization |
|---|---|---|
| Fenite-carbonatite-granite pegmatite | 1070–1040 | Calcite-apatite veins, carbonatite, U-Th, Mo, rare element pegmatites, feldspar, mica, corundum; most type 8 metallic deposits associated with this suite |
| **Skootamatta Suite**<br>Monzonite-diorite | 1090–1075 | No known mineralization, some plutons (e.g., Mount Moriah Syenite) have been evaluated for building stone potential |
| **Gananoque Suite**<br>Syenite-monzonite | 1180–1160 | No known mineralization; some plutons are exploited for building stone (e.g., Lyndhurst and Battersea granites) |
| **Methuen Suite**<br>Alaskite | 1250–1240 | Contact metasomatic iron deposits, contact metamorphic talc, wollastonite, dolomite, tremolite deposits; high fluorine bodies have associated Ba-F veins and potential for Nb-Ta-Cu deposits |
| **Lavant Suite**<br>Diorite-gabbro | 1250–1240 | Fe-Ti magmatic oxide deposits, contact metasomatic iron deposits, contact metasomatic talc, wollastonite, tremolite deposits, contact metamorphic stratiform zinc deposits |
| **Elzevir Suite**<br>Tonalite | 1280–1270 | No known mineralization, local stone potential |
| **Killer Creek Suite**<br>Gabbro-anorthositic gabbros ("Older Gabbros") | 1300–1275 | Cu-Ni magmatic sulphide deposits, possible magmatic Cr-Ti and PGE deposits |
| **Anorthosite Suite** | 1280–1250 | Alumina source |
| **Nepheline Syenite Suite** | 1280–1250 | Alumina source (ceramic use), magnetite concentrations |
| **Dysart Suite**<br>Early tonalites | 1370–1340 | No known mineralization |

Jackson and Fyon, this volume). The abundance of migmatitic quartzofeldspathic gneisses (Red Cedar Lake Gneiss; Lumbers 1978; Easton 1988b) reflects this metasedimentary derivation, and this segment might be expected to have the same mineral potential as other metasedimentary terranes of the Superior Province, as described by Fyon, Breaks et al. (this volume). The Red Cedar Lake Gneiss has a very attractive appearance, however, a high biotite content and considerable jointing render it a poor stone prospect. The lack of orthogneisses in this segment of the Grenville Front Tectonic Zone makes it an unfavourable region for stone prospects.

## Nipissing Terrane

The Nipissing Terrane is underlain by large areas of Archean age (Krogh 1989; Dickin and McNutt 1990) migmatitic gneisses. The polymetamorphic character of these rocks prevents any realistic evaluation of their mineral potential, although it is likely to be limited. Supracrustal rocks in the French River area may be Paleoproterozoic to Mesoproterozoic in age, and include metamorphosed quartz arenite, quartzofeldspathic and amphibolitic gneisses. This region may have the best potential for mineralization, although a reasonable estimate of potential commodities in the area is hampered by the lack of previous prospecting activity. Placer gold concentrations in the metamorphosed quartz arenites and base metals associated with the quartzofeldspathic gneisses and amphibolites might be expected. Possible iron formation also occurs in Hendrie Township within calc-silicate gneiss (Lumbers 1975). Specularite-rich layers occur in metamorphosed quartz arenites in Cosby and Delamere townships in the French River area (C. Marmont, Ontario Ministry of Northern Development and Mines, Dorset, written communication, 1991).

Several metaplutonic suites occur in the Nipissing Terrane, and their building stone potential has been evaluated by Lacey (1989). These include the Cosby Granite, the Rutter Granite and the St. Charles Anorthosite. Nepheline syenite of the Bigwood Alkalic Complex (Hewitt 1961) is a potential source of nepheline syenite (for use as an industrial mineral), however, it is not as pure or as large as the Blue Mountain Nepheline Syenite currently being exploited in the Central Metasedimentary Belt. Several Neoproterozoic mafic to ultramafic intrusions occur in the terrane and have associated pyrite and chalcopyrite mineralization (*see* Easton, this volume).

The area is cut by Ottawa–Bonnechere graben faults and several Neoproterozoic to Early Cambrian alkalic complexes hosting iron, niobium and thorium mineralization are present near Lake Nipissing (Currie 1976; *see* Sage, this volume; *see also* Easton, this volume). Fenitization aureoles are associated with the complexes (Lumbers 1971a; Currie 1976). Uranian pyrochlore, iron-titanium minerals and minor sulphide mineralization occur in some of the alkalic complexes (Lumbers 1971a). In the Iron Island Alkalic

Complex, massive to disseminated iron-titanium minerals and pyrite, pyrrhotite and minor chalcopyrite occur in all intrusive phases; but are concentrated in alkalic pyroxenite and carbonatite (Lumbers 1971a). Uranian pyrochlore is concentrated in the silicocarbonate and ijolitic rocks (Lumbers 1971a). Five deposits of uranian pyrochlore occur in the Manitou Island Alkalic Complex (Rowe 1958; Currie 1976), concentrated in the aegirine-potassic feldspar fenite zone. The Allen Lake Carbonatite also contains anomalous concentrations of iron, niobium, thorium and rare earth elements (Ford et al. 1988).

## Tomiko Terrane

The Tomiko Terrane consists of Mesoproterozoic supracrustal rocks dominated by paragneisses, including the Mattawa Quartzite. Lumbers (1971b, 1971c) mapped the terrane in detail, and identified several areas of quartz-magnetite iron formation, which is rare elsewhere in the Central Gneiss Belt. The abundance of supracrustal rocks in this terrane suggests a high mineral potential. Several stone quarries are present in the northern part of the Tomiko Terrane (Owsiacki et al. 1990). Kyanite (Owsiacki et al. 1989) and vermiculite (Lumbers 1971c) also occur.

## Algonquin Terrane

The best inferences on the metallogenic potential of the Algonquin Terrane come from an examination of mineralization types through time and space (e.g., Hutchinson 1981a, 1981b; Veizer et al. 1989). Most Algonquin Terrane rocks were deposited during the Mesoproterozoic to Neoproterozoic (ca. 1.7 to 0.9 Ga) phase of stable cratons (terminology of Veizer et al. 1989). Significant ore deposits in this period are mainly exogenic, primarily unconformity-related uranium, manganese and copper deposits, and are generally found in cratonic sedimentary cover sequences. As outlined in Easton (this volume), the rock types present in some domains within the Algonquin Terrane (e.g., Britt, Go Home, Kiosk and Powassan domains) have significant metasedimentary components, broadly consistent with the dominant mineral depositional environments existing during this period.

Mineral deposits found in high-grade metamorphic terranes include the Broken Hill lead-zinc deposits of Australia, which occur at upper amphibolite to granulite facies, and granulite facies graphite deposits of the Sri Lankan shield. There is considerable debate over the setting of the Broken Hill deposits. Some workers favour a volcaniclastic depositional regime with varied amounts of associated felsic pyroclastic rocks and basic intrusive and extrusive rocks (e.g., Barnes and Willis 1989). Other workers favour a cratonic sedimentary basin, perhaps associated with localized rifting, and basinal dewatering to form the ores (e.g., Haydon and McConachy 1987, 1989; Wright et al. 1987). There is reasonable potential for Broken Hill-type mineralization within the Algonquin Terrane, as many rock types present in the Algonquin Terrane are similar to those associated with the Broken Hill ores, regardless of which model is correct.

The metasedimentary-rich domains of the Algonguin Terrane host significant graphite deposits. The Cal Graphite Mine in Butt Township, discovered in 1917, began producing ore early in 1990 following exploration and development between 1985 and 1990 (Figure 24.7; Garland 1987). Significant graphite deposits have so far only been reported in the Kiosk, Opeongo and Powassan domains in the north-central and northeastern part of the Algonquin Terrane; these include the Laurier, Maria and Ryerson Township prospects (Garland 1987). The regional setting of these deposits and associated rock types are poorly constrained. All graphite deposits occur in granulite facies, quartzose and quartzofeldspathic rocks in high-strain zones located either near, but not at, domain boundaries, or in high-strain zones between orthogneiss intrusions. Detailed mapping of the Cal Graphite property by Garland (1991) indicated significant structural control in localizing flake graphite mineralization, and clearly indicated that the deposit occurs within a shear zone. Davidson (1982) noted several graphite occurrences within the Algonquin Terrane, the larger of which were examined by Garland (1987). Exploration for additional graphite deposits should focus on the metasedimentary-rich domains of the Algonquin Terrane, particularly areas containing abundant pelitic and semipelitic gneisses adjacent to major shear zones. Geophysical prospecting methods may be useful in graphite exploration. Marmont (1991) summarized the building stone, flagstone and feldspar resource potential of the western Algonquin Terrane.

## Parry Sound Terrane

The abundance of mafic rocks and carbonate and pelitic gneisses in the Parry Sound Terrane makes this region the most extensively prospected, and the best exploration target in the Central Gneiss Belt. Villard et al. (1984) summarized numerous gold prospects in the area. Most are associated with quartz veins and contain associated chalcopyrite, pyrite and pyrrhotite, and are spatially associated with

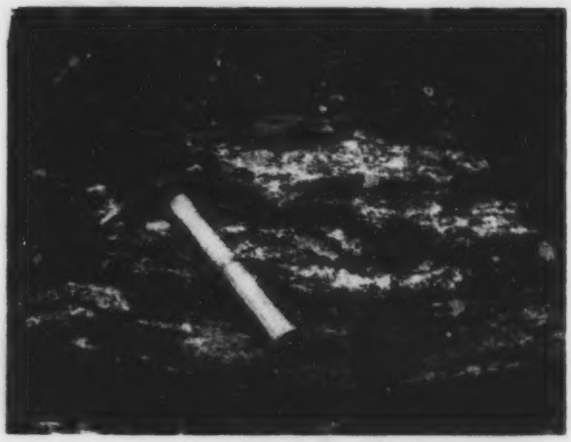


**Figure 24.7.** Coarse-flake graphite ore (beside hammer) at the Cal Graphite graphite mine in Butt Township, Central Gneiss Belt. Hammer handle is 30 cm long.

metamorphosed quartz diorite bodies surrounded by mafic gneisses. Two vein types are documented: 1) copper-gold-bearing veins in gabbroic rocks in mafic gneiss; and 2) copper- nickel- and copper-zinc-bearing veins in the contact zone between felsic and mafic gneiss. Minor past production was achieved from a few deposits (Satterly 1943; Garland and Villard 1990). Small bodies of massive magnetite occur in Lount Township within garnet-hornblende gneiss (Satterly 1956) and may represent disaggregated or disrupted iron formation (C. Marmont, Ontario Ministry of Northern Development and Mines, Dorset, written communication, 1991).

The industrial mineral potential of the Parry Sound Terrane is high, and includes alumina sources (mainly meta-anorthosite; Marmont and Johnston 1987; Marmont 1988a), agricultural lime from marbles (Marmont 1988b), building stone (Marmont 1991) and flagstone (Fouts and Marmont 1989). The best flagstone quarries are located in the Parry Sound shear zone.

# MINERAL DEPOSITS IN THE CENTRAL METASEDIMENTARY BELT

## Metallic Minerals

Metallic mineralization in the Central Metasedimentary Belt includes concentrations of iron, titanium, molybdenum, uranium, thorium, copper, lead, zinc, nickel, gold and silver (Sangster and Bourne 1982; Carter 1984; Carter and Colvine 1985). These metallic minerals are classified into the mineral deposit types listed in Table 24.1. Deposit types 1 to 6 developed prior to, or during, the Elzevirian Orogeny. Epigenetic gold-quartz-carbonate veins (deposit type 7) possibly formed during at least 2 stages, prior to, and immediately following the Elzevirian Orogeny. Deposit type 8 formed after the peak metamorphism of the Ottawan Orogeny. Deposit types 9 and 10 are Phanerozoic in age, but are discussed in this chapter because the mineralization is superimposed upon Grenville rocks.

Economically significant metal production has been attained from the Marmoraton contact metasomatic iron deposit (type 6), the Kingdon lead-bearing vein deposit (type 9), the Cordova and Deloro gold deposits (type 7), the Long Lake carbonate-hosted, stratiform zinc deposit (type 2), several pyrite deposits (type 1, Blithfield, Blakely, Hungerford and Canada mines) and the Madawaska (Faraday) uranium deposit (type 8).

## Importance of Stratigraphic Control

Carter (1984) and Carter and Colvine (1985), in particular, noted the importance of stratigraphy and rock type in controlling the distribution of metallic mineral deposits in the Central Metasedimentary Belt (Figure 24.8 and Table 24.3). This does not mean that all deposits are stratiform or stratabound, rather only that the specific set of circumstances controlling mineralization are strongly influenced by host rocks. For example, the unconformity between the Flinton Group and the Mazinaw Group (Grenville Supergroup) has served as a locus for mineralization. It is still not clear what role the Flinton Group plays in the mineralization (e.g., good cap rock, source of metals, compositional control on mineralizing fluids, structural control related to the depositional environment of the Flinton Group, supergene enrichment of underlying strata during weathering or fluid movement during subsequent metamorphism, some combination of the above), however, the contact between the Flinton Group and the Mazinaw Group can be targeted for exploration for specific mineral deposits which have no direct counterpart elsewhere in the Central Metasedimentary Belt. Thus, an understanding of the lithologic diversity and tectonic history of the various terranes and domains is relevant to mineral exploration in the Central Metasedimentary Belt.

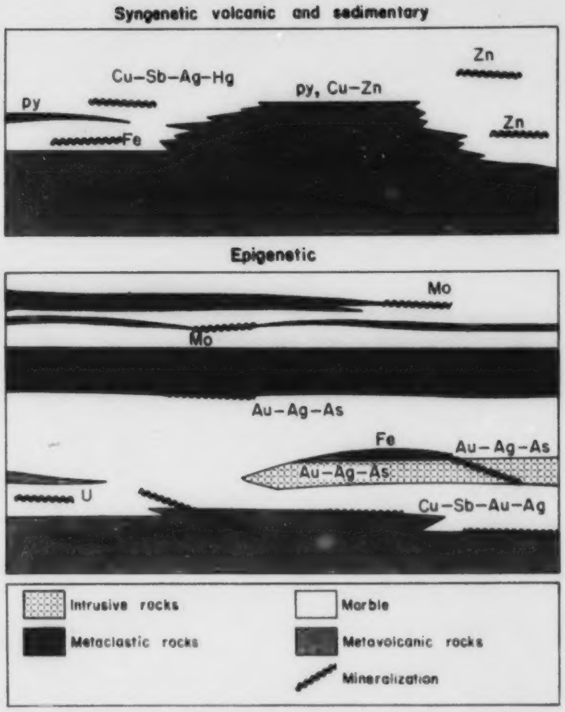


**Figure 24.8.** Relationships between regional stratigraphy and mineralization in the Central Metasedimentary Belt as envisaged by Carter (1984) and Easton et al. (1986).

Figures 24.9a to 24.9f summarize the distribution of types of mineral deposits between the different terranes and domains. Certain deposit types are restricted to specific terranes. Other deposit types, although more widespread, may be of greater size or grade in some terranes or domains than in others. For example, economic and subeconomic stratiform zinc mineralization in dolomite marbles seems to be concentrated in the Bancroft and Sharbot Lake terranes, even though examples of such mineralization occur throughout the Central Metasedimentary Belt. Some deposit types are associated with terrane boundaries. For example, major gold occurrences are more common along the eastern and western margins of the Grimsthorpe Domain, particularly where greenschist facies metavolcanic rocks are juxtaposed against amphibolite facies rocks. Similarly, gold occurrences in the Sharbot Lake Terrane occur along the western margin of the terrane in the Robertson Lake mylonite zone, where the largest deposits are associated with greenschist facies rocks. In these instances, suitable host rocks and specific metamorphic and structural conditions are all responsible for mineralization. Table 24.3 outlines the relation of mineral deposit types to the main plutonic suites present in the Central Metasedimentary Belt. Table 24.4 provides a summary of descriptive mineral deposit studies within the Central Metasedimentary Belt.

Figure 24.10 illustrates the distribution of metallic and industrial mineral deposits in the contact aureole of plutons intrusive into carbonate rocks. The figure is loosely modelled after the Deloro Granite which intrudes greenschist and lower amphibolite facies rocks in the Madoc area. In this case, a variety of factors influence the types and styles of mineralization, such as suitable host lithology, distance from the intrusion, local structural controls, subsequent metamorphic and deposition history, and subsequent uplift and weathering. Thus, although Figure 24.10 provides some guides to exploration, the presence or absence of mineralization in any particular area cannot be predicted without a detailed understanding of the local geology and the regional geologic history. Further, in areas subjected to higher grades of metamorphism, recognition of a contact aureole around a pluton and its role in mineralization may be obscured by metamorphic overprinting and superimposed

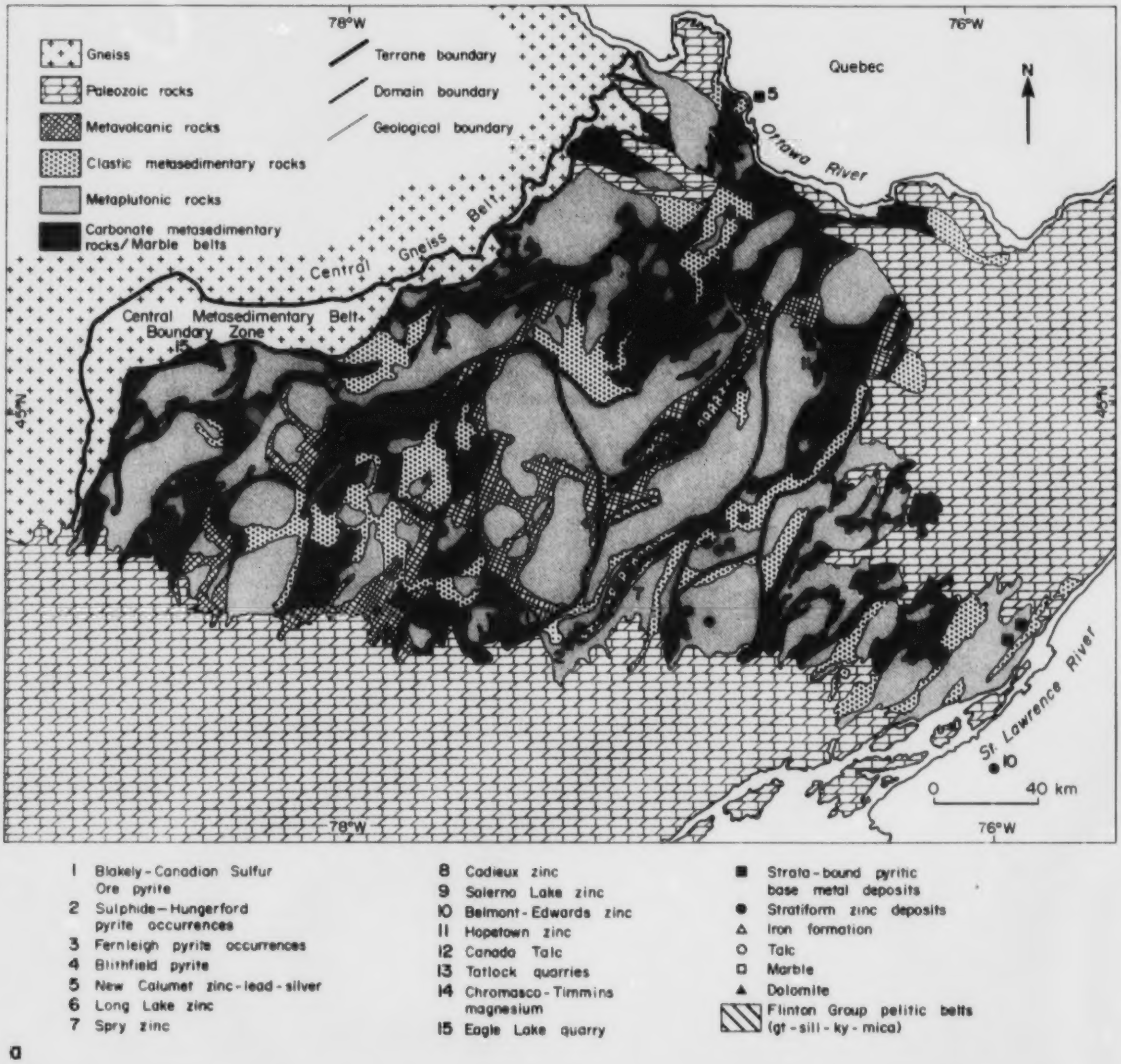

78°W
76°W
Gneiss
Paleozoic rocks
Metavolcanic rocks
Clastic metasedimentary rocks
Metaplutonic rocks
Carbonate metasedimentary rocks/Marble belts
Terrane boundary
Domain boundary
Geological boundary
Quebec
N
5
Ottawa River
Central Gneiss Belt
Central Metasedimentary Belt
Boundary Zone
15
45°N
St. Lawrence River
10
0
40 km
1 Blakely-Canadian Sulfur Ore pyrite
2 Sulphide—Hungerford pyrite occurrences
3 Fernleigh pyrite occurrences
4 Blithfield pyrite
5 New Calumet zinc-lead-silver
6 Long Lake zinc
7 Spry zinc
8 Cadieux zinc
9 Salerno Lake zinc
10 Belmont-Edwards zinc
11 Hopetown zinc
12 Canada Talc
13 Tatlock quarries
14 Chromasco-Timmins magnesium
15 Eagle Lake quarry
■ Strata-bound pyritic base metal deposits
● Stratiform zinc deposits
△ Iron formation
○ Talc
□ Marble
▲ Dolomite
▨ Flinton Group pelitic belts (gt-sill-ky-mica)
a

**Figure 24.9a.** Distribution of syngenetic sulphide deposits in the Central Metasedimentary Belt and significant industrial mineral belts (talc, Flinton pelites, pure carbonates) (*modified from* Carter 1984). Terrane boundaries as shown in Figure 24.6.

deformation. Thus, the utility of this model may be restricted to the lower metamorphic grade parts of the Central Metasedimentary Belt.

## Industrial Minerals

Many workers treat industrial mineral deposits separately from metallic mineral deposits. In the Central Metasedimentary Belt, stratigraphy plays a major role in localizing these deposits as well, and in this chapter, industrial and metallic minerals are discussed together. As shown in Figure 24.10, specific industrial mineral deposits are associated with pluton contact aureoles where suitable host rocks and other conditions are met. The Flinton Group is the major target stratigraphic unit for deposits of sillimanite, kyanite, muscovite and garnet, mainly as it is the only alumina-rich metasedimentary unit within the Central Metasedimentary Belt. Locating this unit is a necessary first condition to finding a suitable industrial mineral prospect for these elements, with secondary factors being purity of the deposit, appropriate grain size, proximity to road network and so on. As a consequence, Figures 24.9a to 24.9f highlight the various industrial mineral prospects of the various terranes. Table 24.4 lists specific commodity studies within the Central Metasedimentary Belt.

Building stone (ashlar, facing, flagstone, ornamental stone) also shows similar stratigraphic controls, but secondary factors such as fracture pattern, proximity to transportation networks, ease of quarrying, and so on are probably more critical factors in determining the marketability of these products. Although many pluton types in the Central Metasedimentary Belt might seem to be suitable and attractive building stones, when all these factors are weighed, few building stone prospects can be identified (*see* LeBaron et al. 1990). Of all the various building stone uses, flagstone prospects are probably the

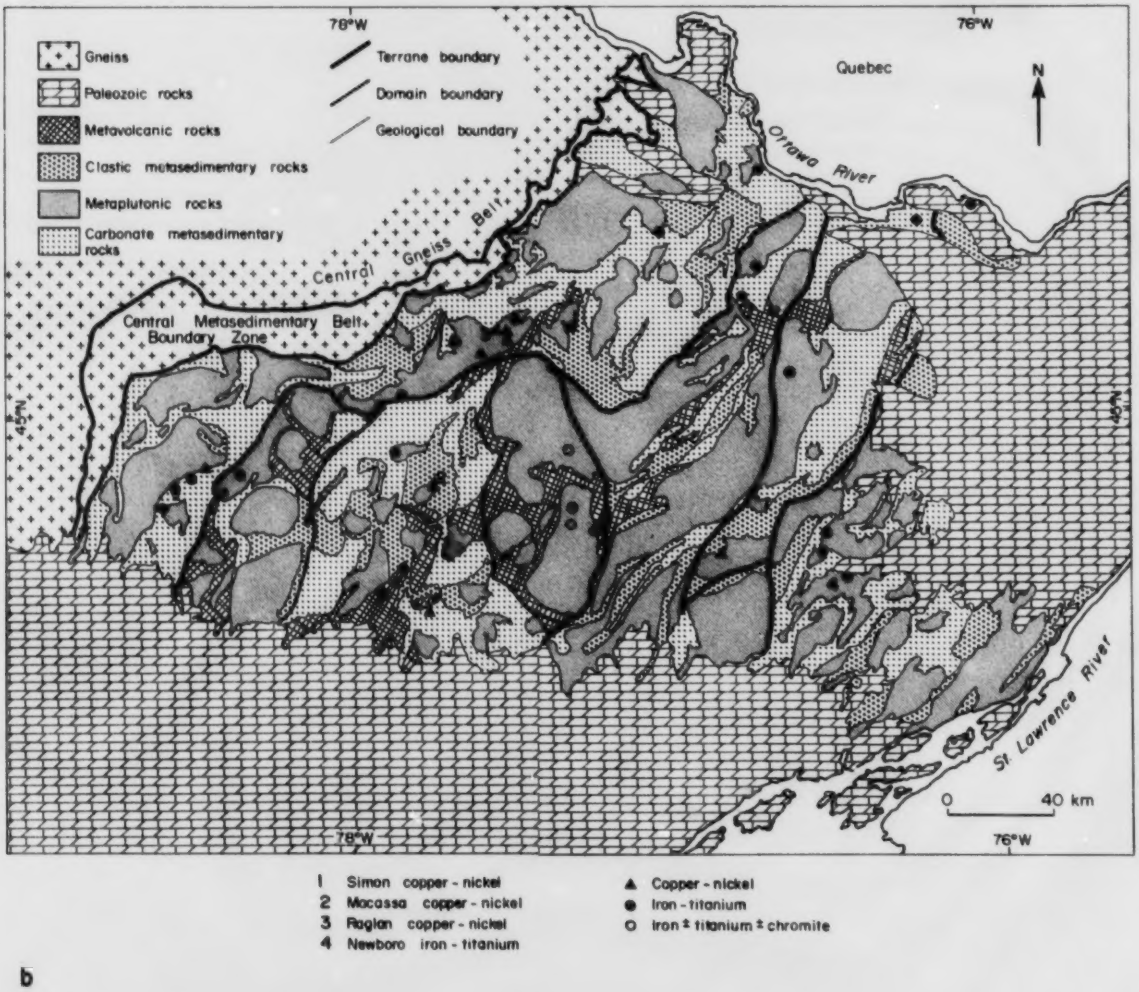

78°W
76°W
Gneiss
Paleozoic rocks
Metavolcanic rocks
Clastic metasedimentary rocks
Metaplutonic rocks
Carbonate metasedimentary rocks
Terrane boundary
Domain boundary
Geological boundary
Quebec
N
Ottawa River
Central Gneiss Belt
Central Metasedimentary Belt Boundary Zone
45°N
St. Lawrence River
0
40 km
1 Simon copper - nickel
2 Macassa copper - nickel
3 Raglan copper - nickel
4 Newboro iron - titanium
▲ Copper - nickel
● Iron - titanium
○ Iron ± titanium ± chromite
b

**Figure 24.9b.** Distribution of copper-nickel and iron-titanium deposits in the Central Metasedimentary Belt (*modified from* Carter 1984). Terrane boundaries as shown in Figure 24.6.

most easily identified on a regional scale, as they are restricted to ductile deformation zones subjected to middle to upper amphibolite facies metamorphism. Thus, domain and terrane boundaries in the Central Gneiss Belt and the Central Metasedimentary Belt Boundary Zone are the primary target areas for flagstone deposits.

Economically significant production of industrial minerals has been attained from the Blue Mountain Nepheline Syenite (ceramic industry), the Chromasco dolomite deposit (used for magnesium metal production), the Tatlock marble quarry (*see* Figure 24.4; fine-ground calcium carbonate products), the Canada Talc Mine (talc; *see* Figure 24.5), graphite (Black Donald and Globe mines), the quarries associated with the 3M plant near Havelock (roofing granules) and numerous building stone quarries (e.g., Figure 24.11). Table 24.4 lists specific commodity studies within the Central Metasedimentary Belt.

## Volcanism, Sedimentation and Mineralization

### SULPHIDE MINERALIZATION

Strata-bound, base metal-bearing, sulphide mineralization (type 1 deposits) occurs in all terranes of the Central Metasedimentary Belt (*see* Figure 24.9a), in 2 general rock associations (Carter and Colvine 1985): 1) diverse successions consisting of coeval mafic and felsic metavolcanic rocks and volcanic-derived, clastic metasedimentary rocks, including units of dolomitic marble (e.g., Blithfield pyrite mine, Mazinaw Terrane; Deer Lake and Caverly deposits, Belmont Domain, Elzevir Terrane; *see* Figure 24.9a; Carter 1984); and 2) metasedimentary successions, which may be spatially associated with older metavolcanic rocks, but which lack coeval metavolcanic rocks (e.g., Sulphide–Hungerford deposits, Clare River area, Mazinaw Terrane;

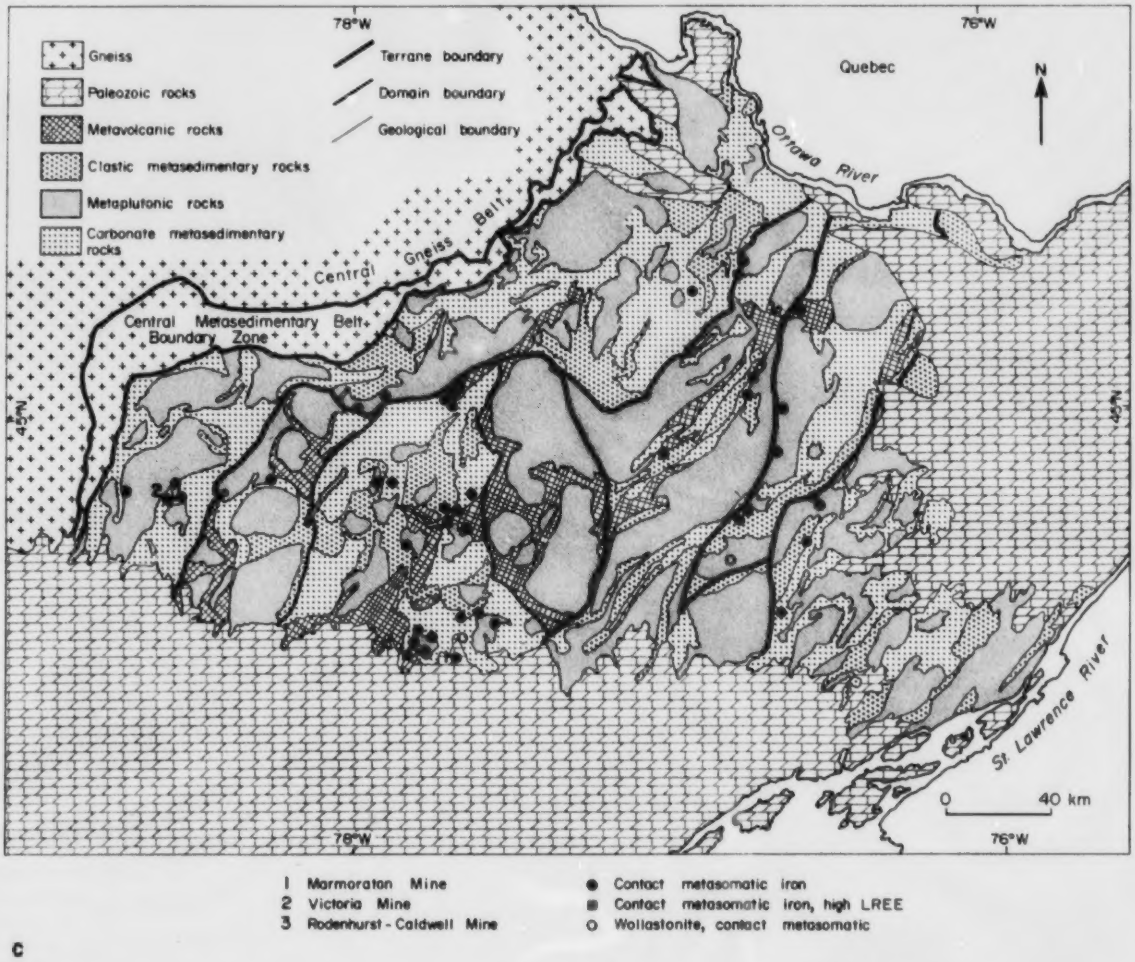

Gneiss
Paleozoic rocks
Metavolcanic rocks
Clastic metasedimentary rocks
Metaplutonic rocks
Carbonate metasedimentary rocks
Terrane boundary
Domain boundary
Geological boundary
78°W
76°W
45°N
45°N
Quebec
N
Ottawa River
Central Gneiss Belt
Central Metasedimentary Belt Boundary Zone
St. Lawrence River
0
40 km
78°W
76°W
1 Marmoraton Mine
2 Victoria Mine
3 Rodenhurst - Caldwell Mine
● Contact metasomatic iron
■ Contact metasomatic iron, high LREE
○ Wollastonite, contact metasomatic
c

**Figure 24.9c.** Distribution of contact metasomatic iron deposits in the Central Metasedimentary Belt and contact metasomatic industrial minerals (wollastonite, talc) (*modified from* Carter 1984). Terrane boundaries as shown in Figure 24.6.

Furnace Falls area, Bancroft Terrane; *see* Figure 24.9a; Carter 1984).

Strata-bound mineralization in both associations generally consists of pyrite and pyrrhotite, with minor chalcopyrite, sphalerite and galena (Carter and Colvine 1985). The sulphide minerals are disseminated or occur as massive lenses, pods or zones, generally parallel to the local layering or foliation (Carter 1984) and hosted in carbonaceous and sulphide-rich shales at the top of metavolcanic successions (Figure 24.12). The sulphide-bearing units may be overlain by dolomitic and calcitic marbles. Sulphide content in the deposits ranges up to 20% (Carter 1984) and several deposits in both associations were mined as sources of sulphur in the late nineteenth and early twentieth centuries. These include the Blakely Mine, Elzevir Terrane (Wilson 1921; Carter 1984), Blithfield Mine, Mazinaw Terrane (Wilson 1921; Carter et al. 1980; Easton 1988a) and Hungerford and Canada mines, Mazinaw Terrane (Wilson 1921; Bright 1986). Most of the massive pyritic deposits are not associated with anomalous concentrations of base and precious metals (Carter 1984).

## Volcanic-Associated Deposits

Many of the strata-bound sulphide occurrences are spatially associated with metavolcanic rocks and some occur at, or proximal to, volcanic centres (Carter 1984; Carter and Colvine 1985). Most of the sulphide concentrations were deposited in carbonaceous and sulphide-bearing shales (*see* Figure 24.11), or in dolomitic marbles, at the top of volcanic successions. The shales accumulated in an anoxic depositional environment, following a volcanic episode. The deposits are interlayered with, or are overlain by, marbles and volcaniclastic metasedimentary rocks which accumulated in an oxygenated environment. The

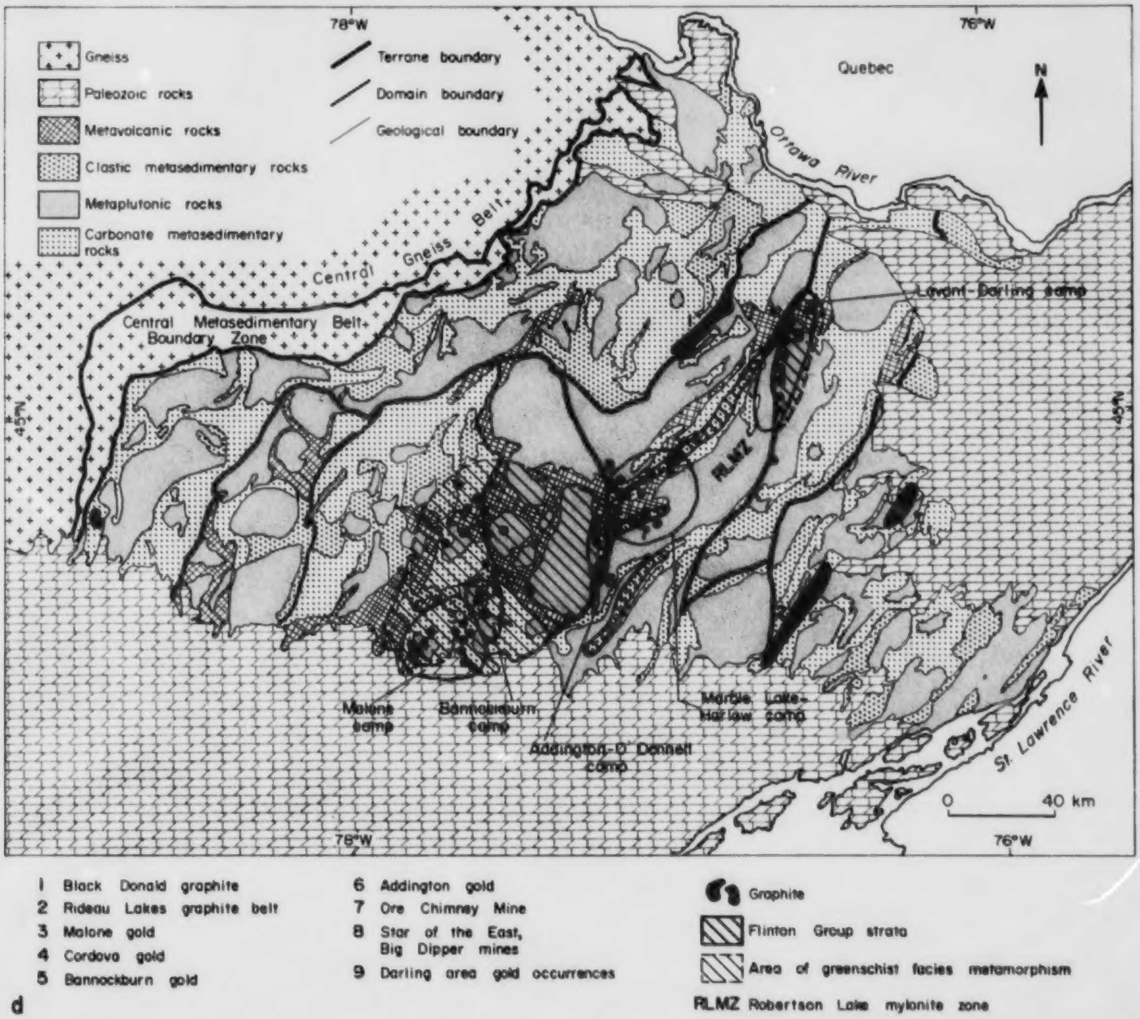

Gneiss
Paleozoic rocks
Metavolcanic rocks
Clastic metasedimentary rocks
Metaplutonic rocks
Carbonate metasedimentary rocks
Terrane boundary
Domain boundary
Geological boundary
78°W
76°W
45°N
Quebec
N
Ottawa River
Central Gneiss Belt
Central Metasedimentary Belt Boundary Zone
Lavant-Darling camp
RLMZ
Malone camp
Bannockburn camp
Addington-O'Donnell camp
Marble Lake-Harlow camp
St. Lawrence River
0
40 km
1 Black Donald graphite
2 Rideau Lakes graphite belt
3 Malone gold
4 Cordova gold
5 Bannockburn gold
6 Addington gold
7 Ore Chimney Mine
8 Star of the East, Big Dipper mines
9 Darling area gold occurrences
Graphite
Flinton Group strata
Area of greenschist facies metamorphism
RLMZ Robertson Lake mylonite zone
d

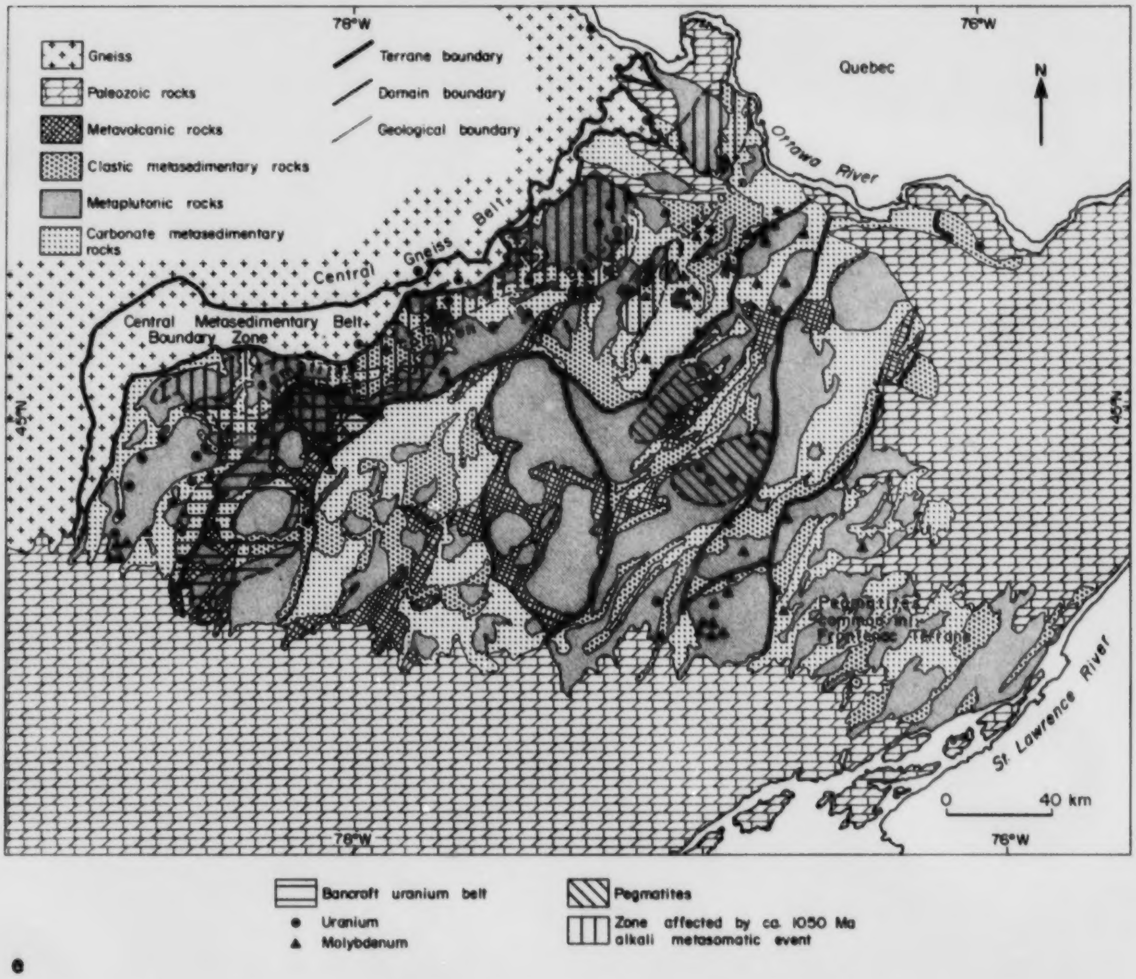
**Figure 24.9d.** Distribution of gold-quartz vein deposits and graphite belts in the Central Metasedimentary Belt (*modified from* Carter 1984). Terrane boundaries as shown in Figure 24.6.

strata-bound, stratiform base metal-bearing sulphide deposits are interpreted to be syngenetic deposits (Sangster 1970; Carter 1984; Carter and Colvine 1985). Supracrustal rocks in the Central Metasedimentary Belt were deposited in several different depositional environments. The sulphide occurrences are not restricted to any specific setting, as outlined below.

The strata-bound pyritic occurrences in the northern Sharbot Lake Terrane and in the Grimsthorpe Domain are associated with ca. 1290 to 1270 million-year-old tholeiitic basalts interpreted to have been erupted as mafic, oceanic plains or oceanic shields or in volcaniclastic sediments derived from erosion of these mafic edifices (*see* Easton, this volume). Although the bulk of the metabasalts may have been deposited in deep water, the sulphide occurrences are associated with volcanic centres that may have reached moderate, or even shallow, water depths, and occur near the volcanic-sediment interface.

The strata-bound pyritic occurrences in the Flower Station and Marble Lake areas in the Mazinaw Terrane are hosted within a younger (1255 to 1245 million-year-old), calc-alkalic basalt-andesite-dacite-rhyolite succession. Again, the occurrences occur near the volcanic-sediment interface, and were likely deposited in relatively shallow water environments. Pyrite occurrences in the Belmont Lake area, Elzevir Terrane, occur near the top of a bimodal tholeiitic basalt-dacite-rhyolite succession dated at 1248 to 1254 Ma (Davis and Bartlett 1988). These younger volcanic sequences may have been erupted as ocean islands or as sea-facing volcanic edifices along the edge of small landmasses.

The ca. 1255 to 1245 million-year-old bimodal tholeiitic basalt-dacite-rhyolite successions have been assigned to back-arc or marginal basin settings on the basis of geochemistry (e.g., Smith and Holm 1987, 1990a, 1990b; Bartlett 1983). Most Archean and Phanerozoic



Gneiss
Paleozoic rocks
Metavolcanic rocks
Clastic metasedimentary rocks
Metaplutonic rocks
Carbonate metasedimentary rocks
Terrane boundary
Domain boundary
Geological boundary
78°W
76°W
Quebec
N
Ottawa River
Central Gneiss Belt
Central Metasedimentary Belt Boundary Zone
45°N
Pegmatites common in Frontenac Terrane
St. Lawrence River
0
40 km
Bancroft uranium belt
Uranium
Molybdenum
Pegmatites
Zone affected by ca. 1050 Ma alkali metasomatic event
e

**Figure 24.9e.** Distribution of uranium and molybdenum deposits within the Central Metasedimentary Belt (*modified from* Carter 1984). Terrane boundaries as shown in Figure 24.6.

volcanic-associated, massive, base metal sulphide (VMS) deposits occur in similar extensional settings (e.g., Franklin 1986), and on this basis alone, VMS mineral potential is high. Further, it has been suggested that VMS mineralization is more likely to occur if icelandites and high-silica rhyolites are present (e.g., Lesher et al. 1986; Barrie and Shirey 1991; Macdonald et al. 1990). Although icelandites have not been positively identified from the Central Metasedimentary Belt, data presented by Smith and Holm (1987, 1990a) suggests their presence.

The New Calumet zinc-lead-silver and Montauban zinc-lead and gold-silver deposits in Quebec (Sangster 1970; Jourdain et al. 1987; Villeneuve 1987) represent mineralization of this type (Carter 1984; Carter and Colvine 1985). The New Calumet deposit, located on New Calumet Island in the Ottawa River, northwest of Renfrew, occurs near the metavolcanic-metasediment interface and is associated with a probable satellite volcanic centre. It is not clear which terrane the deposit is in, although it is most likely in the extreme northeastern part of the Mazinaw Terrane. The Montauban deposit is hosted in a small belt of supracrustal rocks in central Quebec not directly contiguous with the Central Metasedimentary Belt, and recent mapping by L. Nadeau (Geological Survey of Canada, Quebec City, personal communication, 1992) suggests that this deposit is part of a metasedimentary succession significantly older than the Central Metasedimentary Belt.

## Sediment-Associated Deposits

Occurrences of the second association are associated with siliceous clastic metasedimentary rocks. Two major types are present. The first is associated with metasedimentary rocks of the Flinton Group. The second type is commonly found in Grenville or Frontenac supergroup rocks, near the Paleozoic unconformity (e.g., Furnace Falls and Elizabethtown areas, Figure 24.9a; Carter 1984). In the latter type, weathering prior to deposition of the Paleozoic rocks may have been an important factor in the genesis or enrichment of these deposits.

The Clyde Forks copper-antimony-silver-mercury-barium occurrence (Nikols 1972; Carter et al. 1980) represents a unique deposit type. It is a conformable deposit hosted in dolomitic marbles and quartzofeldspathic metasedimentary rocks of the Mazinaw Group, close to the trace of the Clyde River fault. Nikols (1972) proposed that mineralization resulted from hydrothermal springs discharging

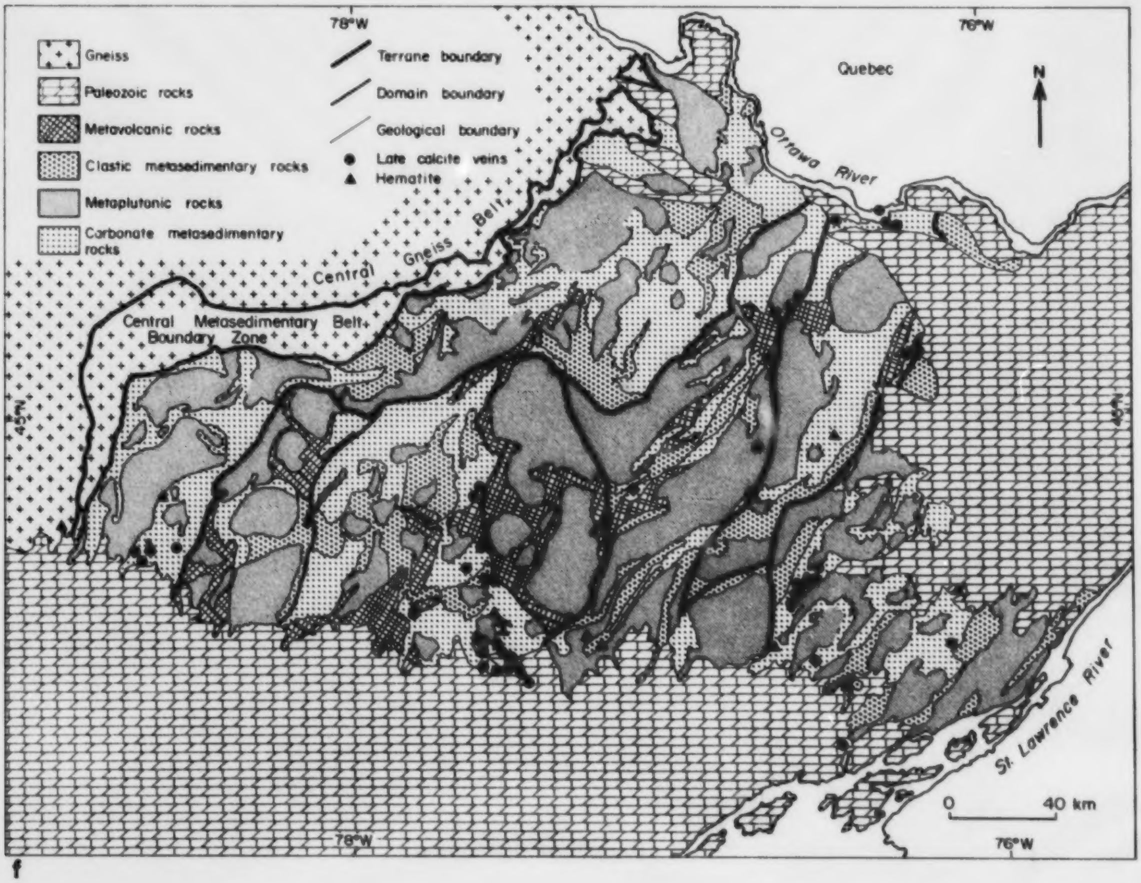

78°W
76°W
Gneiss
Paleozoic rocks
Metavolcanic rocks
Clastic metasedimentary rocks
Metaplutonic rocks
Carbonate metasedimentary rocks
Terrane boundary
Domain boundary
Geological boundary
Late calcite veins
Hematite
Quebec
N
Ottawa River
Central Gneiss Belt
Central Metasedimentary Belt Boundary Zone
45°N
45°N
St. Lawrence River
0
40 km
78°W
76°W
f

**Figure 24.9f.** Distribution of post-Ordovician vein deposits and hematite deposits in the Central Metasedimentary Belt (*modified from* Carter 1984). Terrane boundaries as shown in Figure 24.6.

**Table 24.4.** Summary of descriptive studies of mineral deposits within the Grenville Province, Ontario.

**Metallic Minerals**

Type 1, 2, 3, 4, 5, 6 and 7 deposits: Sangster 1970; Carter et al. 1980; Carter 1984; Malzack et al. 1985

Type 3 deposits: Gross 1967; Carter 1984

Type 5 and 6 deposits: Rose 1958

Type 7 deposits: Sangster 1970; Carter et al. 1980; Carter 1981; Papertzian 1984; Malzack et al. 1985; LeBaron 1991

Type 8 deposits: Satterly 1957; Hewitt 1967a, 1967b; Karvinen 1973; Masson and Gordon 1981; Gordon et al. 1981; Lentz 1991

Type 9 and 10 deposits: Guillet 1963, 1964; Sangster and Bourne 1982; Carter 1984

**Industrial Minerals**

Barite: Guillet 1963

Graphite: Hewitt 1965; MacKinnon and LeBaron 1992

Marble: Goudge 1938; Hewitt 1964; Storey and Vos 1981a; Grant et al. 1989; LeBaron and MacKinnon 1990

Pegmatites: Goad 1990

Building Stone: LeBaron et al. 1989, 1990

Talc: Hewitt 1972; LeBaron and van Haaften 1989

Vermiculite: Guillet 1962; Papertzian and Kingston 1982; MacKinnon et al. 1990

Wollastonite: MacKinnon 1990

Miscellaneous Studies: Storey and Vos (1981b) reviewed industrial mineral prospects in Renfrew County

into a carbonate basin. Flinton Group rocks are present about 1 km east of the deposit. It is not known if hydrothermal activity related to Flinton Group deposition played a role in the genesis of this deposit.

## FLINTON GROUP DEPOSITS

Mineralization is found in 4 settings within metasedimentary rocks of the Flinton Group (Mazinaw Terrane). i) In the Marble Lake area, black pyritic-graphitic schists of the Myer Cave Formation, and locally the Fernleigh Formation contain stratiform pyrite-chalcopyrite-sphalerite mineralization (Figure 24.13). In the Clare River Synform, similar mineralization has been described by Bright (1986) from the Kaladar Mines Limited property and the Amiko mica property. ii) In the Clare River Synform, pyrite-pyrrhotite deposits are closely associated with basal muscovite schists of the Flinton Group (Bright 1986). Examples include the Canada and Hungerford mines near Sulphide (Bright 1986) and the Ontario Sulphur and Donahue Creek deposits (Carter 1984). iii) Bright (1986) reported strata-bound zinc mineralization with minor pyrite and chalcopyrite in dolomitic marbles in the Clare River area, possibly equivalent to the Myer Cave Formation. Examples include the Spry zinc occurrence (Carter 1984) and possibly the Glenshire property (Carter 1984; Bright 1986). iv) Best known in both areas are crosscutting sulphide-quartz veins in the dolomite horizon of the Myer Cave Formation containing pyrite-chalcopyrite-sphalerite-galena mineralization. Examples include the Barrie Syndicate and Mazinaw base metal occurrences (Moore and Morton 1986) and the Ganda Silver Mines occurrence (Pauk 1987). This mineralization is probably the result of remobilization of *kupferschiefer*-type copper-zinc mineralization related to the original deposition of the Flinton Group and represented by types i and iii.

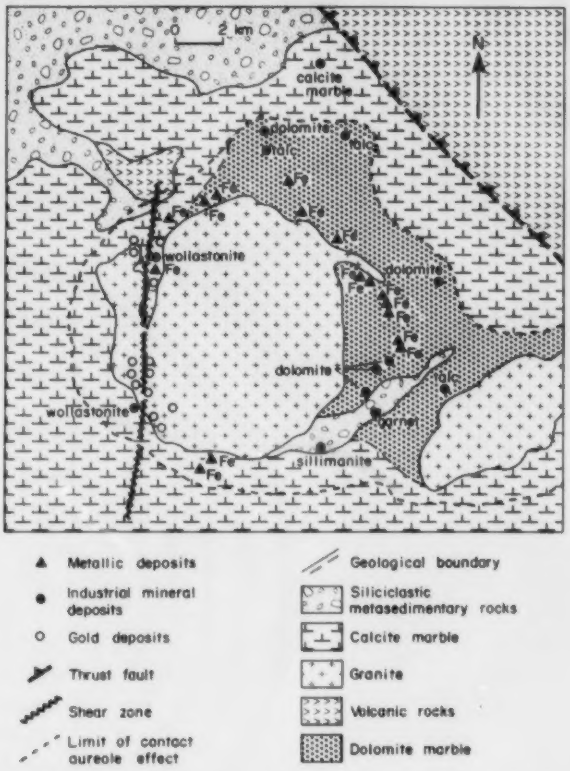


**Figure 24.10.** Sketch map showing metallic and industrial minerals found in contact aureoles around gabbro-diorite and alaskite suite plutons.

**Figure 24.11.** Stone quarry in the Battersea Granite, Frontenac Terrane, operated by Rock of Ages Limited.

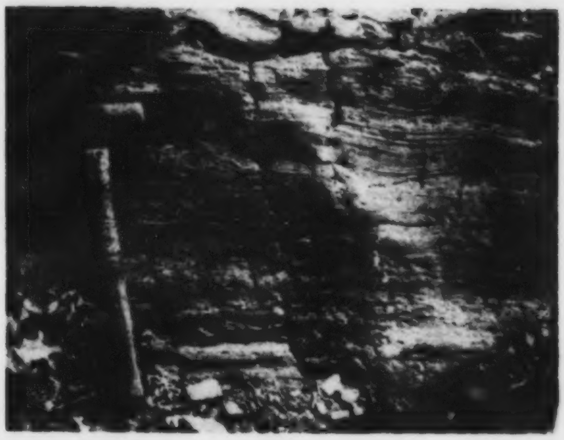


**Figure 24.12.** Pyrite-rich black shale at the top of a mafic metavolcanic succession within the Sharbot Lake Terrane. Sulphidic shales such as this are common near the top of metavolcanic sequences within the Elzevir, Mazinaw and Sharbot Lake terranes of the Central Metasedimentary Belt. Hammer handle is 30 cm long.

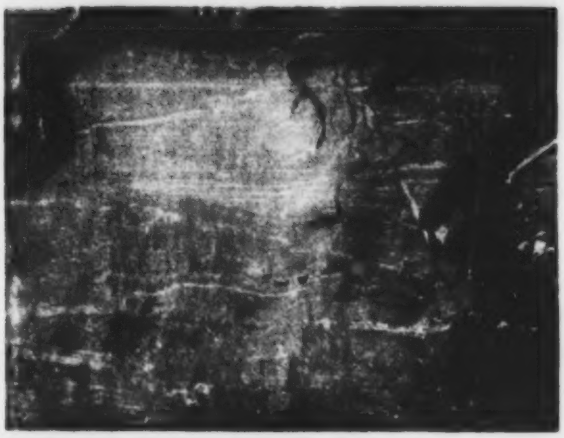


**Figure 24.13.** Pyrite-pyrrhotite-chalcopyrite-rich black shale of the Myer Cave Formation, Flinton Group, Mazinaw Terrane. The Flinton Group succession is some 100 million years younger than the black shale shown in Figure 24.12, and is not associated with volcanism.

## DISCUSSION

The tectonic setting of sedimentary rock-hosted stratabound pyritic occurrences (e.g., Marble Lake and Clare River, Mazinaw Terrane; Furnace Falls, Bancroft Terrane; Elizabethtown, Frontenac Terrane) is difficult to infer. Occurrences in the Furnace Falls and Elizabethtown areas were probably deposited in a platform-slope environment. Occurrences in the Myer Cave Formation of the Flinton Group were probably deposited in a fluvial, lacustrine or shallow marine setting, in an extensional environment (Moore and Thompson 1980; Easton and Ford 1991). The Zambian copper belt (e.g., Lefebvre 1989) is similar in many respects to the postulated environment for Flinton Group deposition and mineralization both within, and adjacent to, the Flinton Group. The low sulphur isotope content of the Spry zinc deposit (2.6‰; Sangster 1991) is consistent with derivation from basinal brines that flowed through adjacent metavolcanic strata (Sangster 1991). Lead is more abundant in Mazinaw Terrane sulphide occurrences than elsewhere in the Central Metasedimentary Belt, perhaps a reflection of the presence of a basement complex of tonalite and gabbroic rocks within the Mazinaw Terrane. Although this basement is only slightly older than the overlying supracrustal rocks, it may have served as a source of lead.

## ZINC MINERALIZATION

Stratiform, carbonate-hosted zinc mineralization (type 2 deposits) constitutes one of the more economically significant deposit types in the Grenville Province in Ontario. Production from the Long Lake Mine of 94 631 short tons of ore averaged 11.6% Zn between March 1973 and December 31, 1974 (Carter 1984). In addition, the Balmat–Edwards district in the Adirondack Lowlands in New York State had, as of 1980, produced 25 million tons of ore containing about 9% Zn and 0.5% Pb (Lea and Dill 1968; deLorraine and Dill 1982). Zinc deposits and occurrences in southeastern Ontario and southwestern Quebec (Figure 24.14) exhibit a variety of geological relationships, but have several common features: 1) deposits occur within carbonate metasedimentary rocks in mixed carbonate-clastic meta-sedimentary successions; 2) deposits are commonly associated with siliceous zones in marbles; 3) zinc-rich sulphide lenses are conformable with layering in the host rocks; 4) sulphide lenses consist principally of sphalerite and pyrite, with a minor tenor of lead and silver; 5) in general, the deposits do not occur in proximity to successions of metavolcanic rock; and 6) deposits have been metamorphosed, with metamorphic mineral assemblages similar to their host rocks (Carter 1984; Carter and Colvine 1985).

In southeastern Ontario, the zinc deposits are widely distributed and occur in the Frontenac, Sharbot Lake, Mazinaw and Bancroft terranes (*see* Figures 24.9a and 24.14), with the larger deposits located in the Bancroft and Sharbot Lake terranes. No deposits are known to occur in the Elzevir Terrane. Host rock associations (Table 24.5) are modified from those recognized by Sangster (1991), and include: 1) calc-silicate and dolomite, spatially associated with metavolcanic rocks; 2) thick units of dolomite in calcitic marble with no spatially associated metavolcanic rocks; and 3) massive calcite marble with minor tremolitic and olivine-rich dolomites spatially associated with gabbroic intrusions. Table 24.5 lists the significant features of these deposit types.

The zinc deposits are apparently restricted to carbonate-rich metasedimentary and siliceous clastic metasedimentary rock successions, especially the shallow to moderate water depth platform and platform-slope environments (Carter 1984; Carter and Colvine 1985; *see* Easton, this volume). Stromatolites are closely associated with many of these deposits (*see* Table 24.5), particularly those hosted in siliceous dolomite units, which implies relatively shallow water depositional environments for the host rocks.

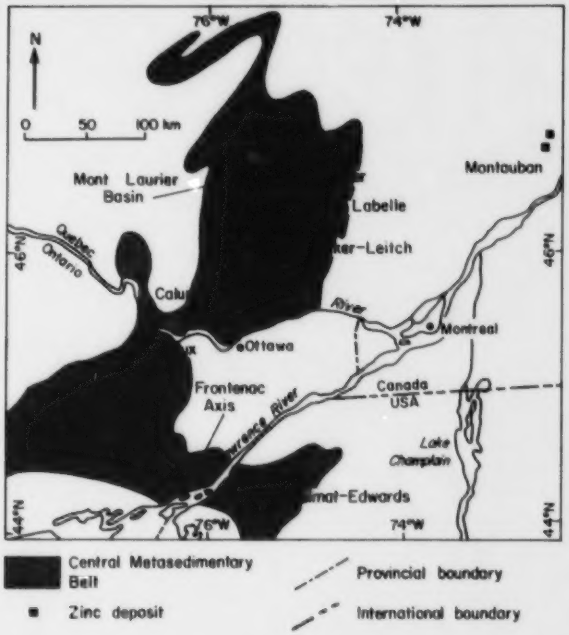


**Figure 24.14.** Distribution of major zinc deposits in the Central Metasedimentary Belt in Ontario and southwestern Quebec. Most deposits are hosted in platformal environments in the Bancroft and Frontenac terranes and their extensions into Quebec (*modified from* Gauthier and Brown 1980).

Such settings likely contributed to early diagenetic dolomitization of these carbonates forming highly permeable host rocks for later mineralizing fluids. Most of the zinc deposits are remote from, and stratigraphically above, volcanic successions. This indicates that direct involvement of volcanism was not important in the formation of the zinc deposits (Sangster 1991).

All of these deposits are interpreted to have formed as stratiform, synsedimentary deposits (e.g., Gauthier and Brown 1980; Lea and Dill 1968; Solomon 1963; deLorraine and Dill 1982; Carter 1984; Carter and Colvine 1985). The mineralizing fluids were likely basinal brines with sulphur derived from adjacent strata, as sulphur isotope values (*see* Table 24.5) vary according to the nature of the host rock succession (Sangster 1991). The Balmat and possibly Salerno deposits may have formed in a tidal mud flat, or sabkha-type environment (e.g., Foose 1980; Carter 1984), based on the presence of anhydrite, tourmalinites and stromatolitic dolomites (Figures 24.15 and 24.16). However, the precipitation of anhydrite and dolomite from epigenetic hydrothermal fluids cannot be ruled out. Although some characteristics of the mineralization and geological setting are similar to those displayed by Mississippi Valley-type zinc deposits, the general setting is also analogous to sediment-associated, exhalative (SEDEX), massive sulphide mineralization (e.g., Krebs 1981; Gustafson and Williams 1981).

Zinc may have precipitated in several tectonic settings. In the Bancroft Terrane, stromatolitic, dolomitic and calcitic marble occur in a succession that also contains arenites, which suggests a shallow water, nearshore, platformal environment. Dolomitization of calcitic marbles may have taken place during diagenesis; sulphide mineralization could have been introduced into the carbonate rocks during the diagenetic, dolomitization event, or shortly afterward. Sulphide minerals in the Salerno deposit have heavy sulphur isotope values (greater than 10‰; Sangster 1991), comparable with those of deposits in the Balmat–Edwards area. These sulphur isotope data indicate interaction between metalliferous brines and carbonate rocks, with no input from a volcanic source (Sangster 1991), which is consistent with the observed depositional setting.

The Cadieux deposit in the northeastern part of the Bancroft Terrane may have formed in a different tectonic setting. The deposit occurs within a succession of dolomitic and calcitic marbles that is bounded by siliceous clastic metasedimentary rocks, in proximity to mafic metavolcanic rocks. Sulphur isotope data from the zinc deposit are consistent with a volcanic sulphur component (Sangster 1991). The sphalerite mineralization may have formed in a marginal basin. Alternatively, it may be a more offshore setting than the Salerno deposit, with deposition occurring in a platform-slope setting that was also receiving detritus from a distal volcanic source region to the southeast. Uncertainty associated with the timing relationship between the volcanic and sedimentary rocks, and the exact location of terrane boundaries within this segment in the Central Metasedimentary Belt, precludes a more rigorous assignment of tectonic setting to this deposit. The Clyde River (Hopetown) deposit in the Sharbot Lake Terrane is hosted in calcitic and dolomitic marbles of the Sharbot Lake Group, which were probably deposited in a platform-slope environment, and thus the overall setting may be similar to the Bancroft Terrane deposits. The tectonic setting of the calcite marble-hosted sphalerite deposits in the Long Lake area in the southern part of the Sharbot Lake Terrane is not well constrained, as these deposits are hosted in roof pendants within Lavant Suite gabbroic intrusions and have been subjected to contact and regional metamorphism.

Marbles hosting zinc mineralization in the Frontenac Terrane may have also formed in a shallow water, platform setting. The increased abundance of dolomitic and stromatolitic marble to the southeast and the presence of evaporite near Balmat, New York, are consistent with the shallowing of the platform to the southeast (*see* Easton, this volume). Carbonate and evaporite deposition may have taken place in a sabkha environment, a setting that favours the introduction of base metals and sulphides into the dolomitic succession. If this facies association is significant in controlling mineralization, then the greatest potential for dolomite-hosted zinc mineralization is in the southeastern part of the Frontenac Terrane.

Zinc deposits in the Mazinaw Terrane (e.g., Spry deposit) may have formed syndepositionally in host dolomite units of the Myer Cave Formation and its stratigraphic equivalents within the Flinton Group. Although a shallow marine setting has been proposed for the carbonates (Moore and Thompson 1980), a fluvial-lacustrine depositional

**Table 24.5.** Zinc deposit types, characteristics and inferred depositional environments within the Central Metasedimentary Belt, Grenville Province.

| Type and Host Rock Associations | Example | Terrane | Age Range (Ma) | Sulphide Mineralogy | Environmental Indicators | Sulphur Isotope Data* | Depositional Environment | Metamorphic Grade |
|---|---|---|---|---|---|---|---|---|
| 1. Calc-silicate and dolomite spatially associated with metavolcanic rocks | Cadieux | Bancroft Terrane | ca. 1250–1270 | sp, minor py, gn | stromatolite found near ore | -2.5‰ | platform-slope, shallow to moderate water depths, near volcanic islands | middle amphibolite |
| | New Calumet | Mazinaw Terrane | ca. 1250–1270 | sp, gn, py associated Au, Ag | stromatolites in mine region | 0.7‰ | | upper amphibolite |
| 2. Thick units of clean dolomitic marble and calcite marble with minor calc-silicate rocks | Spry zinc | Flinton Group, Mazinaw Terrane | ca. 1160 | sp, minor gn | stromatolites nearby, sp with stromatolites reported by Bright (1986) | 2.6‰ | lacustrine or shallow marine in an extensional environment | middle amphibolite |
| | Hopetown | Sharbot Lake Terrane | ca. 1250–1300 | sp, minor gn | — | — | shallow water, platform? offshore of volcanic islands | lower amphibolite |
| | Balmat–Edwards | Frontenac Terrane (Adirondack Lowlands) | ca. 1250–1350 | sp, minor gn | stromatolites at mine and in adjacent stratigraphic units, anhydrite in minor, tourmalinites in mine area | ~14‰ | shallow water platform to sabkha | middle amphibolite |
| | Salerno Lake, Silver Mountain, Buller Lake | Bancroft Terrane | ca. 1250–1300 | sp, minor gn | stromatolites near Buller Lake and Silver Mtn. occurrences | 11.4–15.5‰ | shallow water platform | middle to upper amphibolite |
| 3. Massive calcite marble with tremolitic and olivine-rich dolomitic units, commonly roof pendants in gabbroic intrusions | Long Lake, Slave Lake | Sharbot Lake Terrane | ca. 1250–1270 | sp, minor gn | — | 6.5–11.7‰ | shallow water platform? | middle to upper amphibolite (regional) contact metamorphic grade unknown |

* Data from Sangster (1991).

Abbreviations: Ag - silver; Au - gold ; gn - galena; py - pyrite; sp - sphalerite

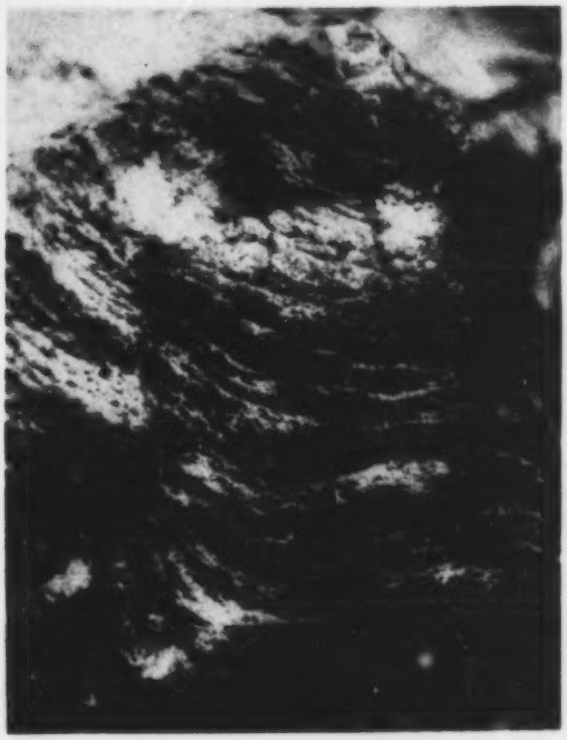


**Figure 24.15.** Stromatolite in dolomite from the Balmat–Edwards no. 2 shaft, Balmat–Edwards zinc district, Adirondack Lowlands (Frontenac Terrane), New York. Field of view is roughly 10 cm across.

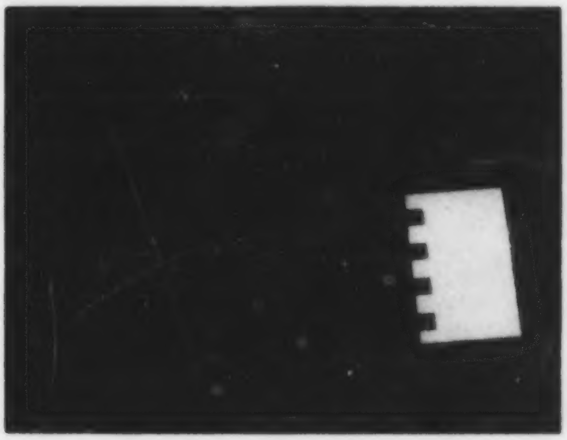


**Figure 24.16.** Tourmalinite beds (dark layers) within the supracrustal succession in the Balmat–Edwards district, Adirondack Lowlands, indicative of a hypersaline depositional environment. Scale bar marked in cm intervals.

setting may be more appropriate (Easton and Ford 1991). In both models, it is suggested that deposition of the Flinton Group occurred in an extensional environment (Moore and Thompson 1980; Easton and Ford 1991). In such an environment, faulting would allow egress of metalliferous hydrothermal fluids into the accumulating clastic and carbonate sediments. Such a syndepositional setting is similar to that in which SEDEX deposits are formed.

### IRON FORMATION

Oxide and sulphide iron formation units (type 3 deposits) are relatively rare in the Central Metasedimentary Belt, but those recognized are texturally and mineralogically similar to oxide-facies Algoma-type (Gross 1967) iron formation units (Carter 1984). The iron formation units have limited strike length and thickness. All occur near, or on the flanks of, volcanic centres within the Belmont Domain (*see* Figure 24.9a); not all volcanic centres have spatially associated iron formation units. Other host rocks can include marble (e.g., Deer Lake occurrence) and felsic metavolcanic rocks (e.g., Limerick Township occurrences; Carter 1984). All are associated with ca. 1250 million-year-old, tholeiitic, bimodal, basalt-dacite rhyolite successions that formed volcanic islands or seaward-facing volcanic edifices along the edge of small landmasses (*see* Easton, this volume).

The oxide facies iron formation units consist of mesobands and discontinuous lenses of fine- to medium-grained magnetite or hematite, alternating with mesobands of fine-grained, recrystallized chert (Carter 1984; Carter and Colvine 1985). Pyrite, chlorite, amphibole and epidote are common accessory minerals. Jasper, hematite and specularite occur in some of the Belmont iron formation units (Carter 1984). The Belmont Lake iron formation is the most extensive, at 15 to 18 m wide and with a strike length of about 400 m (Miller and Knight 1914; Gross 1967). No production has been attained from any of these iron formation units.

### INDUSTRIAL MINERALS

A variety of industrial mineral deposits in the Central Metasedimentary Belt are directly related to volcanism and sedimentation, particularly with respect to the primary composition of the deposits. Examples of deposits showing primary compositional control include relatively pure dolomite and calcite marble horizons. Graphite deposits may also show some original lithologic control, although subsequent deformation and metamorphism are important in concentrating graphite. Most graphite deposits in the Central Metasedimentary Belt occur in granulite facies rocks within the Frontenac Terrane, however, significant deposits are present in upper amphibolite facies rocks within the Mazinaw Terrane (e.g., Black Donald Mine).

## Plutonism and Mineralization

### MAGMATIC COPPER-NICKEL DEPOSITS

Magmatic copper-nickel mineralization (type 4 deposits) is most commonly associated with differentiated mafic to ultramafic intrusions of the Killer Creek Suite that were emplaced into the Central Metasedimentary Belt between 1290 and 1270 Ma. Only a few plutons of this type have so far been delineated (*see* Figure 24.9b).

Four concentrations of copper-nickel mineralization occur within the deformed and metamorphosed Raglan Hills gabbro in the Bancroft Terrane. This is an amphibolite facies, strongly deformed gabbro-anorthosite intrusion, which consists of medium- to coarse-grained, massive to gneissic metagabbro containing accessory titanite, apatite, carbonate, zircon and quartz (Hewitt 1954; Carter 1984). Hornblendite is common around the margins of the intrusion. Pyroxenite and olivine pyroxenite occur in the central part of the intrusion and anorthositic gabbro is also present (Carter 1984). At the Raglan prospect, sulphide mineralization consists of a broad lens of disseminated pyrrhotite, chalcopyrite and pyrite which is 155 m long, 90 m wide and 6 to 15 m thick containing approximately 1% combined copper and nickel (Carter 1984).

The Simon copper-nickel deposit (*see* Figure 24.9b) occurs in an amphibolitic succession composed predominantly of metagabbro, and may be a fragment of the Raglan Hills gabbro. Mineralization consists of disseminated chalcopyrite, pyrrhotite, pyrite and sphalerite contained within a stratiform sulphide lens, similar to that reported elsewhere from the Raglan Hills gabbro. A magnetite-pyrrhotite-pyrite lens is also situated about 100 m structurally below the sulphide lens. The sulphide lens contains up to 250 000 tons of sulphide mineralization containing about 1% Cu and traces of zinc (Carter 1984; Carter and Colvine 1985). Carter (1984) regarded the Simon copper-nickel deposit in the Bancroft Terrane as a potential VMS deposit. Although the Raglan Hills gabbro is lithologically similar to the Killer Creek Suite, preliminary U-Pb zircon geochronology (O. van Breemen, Geological Survey of Canada, personal communication, 1992) suggests an age of 1230 Ma for the complex, contemporaneous with the Lavant Suite.

The Killer Creek Gabbro, within the Grimsthorpe Domain, contains disseminated chalcopyrite; to date no significant mineralization has been uncovered (Easton and Ford, in prep.). Associated with hornblendite pods within this suite are chromite-ilmenite occurrences, containing up to 4000 ppm Cr, locally with PGE enrichments (Pd ~200 ppb, Pt ~20 ppb, Easton and Ford 1991). The largest of these occurrences occurs in Bon Echo Provincial Park. Copper mineralization in the Lingham Lake Gabbro-Diorite Pluton probably is also related to the Killer Creek Suite (Lingham Lake deposit, Lumbers 1969).

Mineralization in the Lavant Suite is exemplified by the Macassa deposit in the Thanet Gabbro and consists of pyrrhotite, pentlandite, chalcopyrite, pyrite, and minor cubanite and sphalerite (Lumbers 1969; Macdonald and Cherry 1988). The mineralized zone is 320 m long, 4 to 34 m wide, and contains an estimated 3.9 million tons grading 0.83% Ni and 0.25% Cu (Carter 1984).

Most of the copper-nickel sulphide occurrences are located at, or near, the margin of the intrusions, consistent with the hypothesis that wall rock assimilation induced sulphide saturation in magmas which do not appear to be sulphide rich (Macdonald and Cherry 1988). The Killer Creek Suite is in part synvolcanic, associated with the older tholeiitic basaltic volcanism that may have been erupted as mafic plains or shields. The Killer Creek Suite gabbros are spatially associated with Elzevir Suite tonalite intrusions which always intrude the gabbros. The Renzy Lake deposit in western Quebec, consisting of disseminated pyrrhotite with pentlandite and chalcopyrite within a metapyroxenite sill (Sangster 1970), may also be of this type. The presence of 2 gabbro suites within the Central Metasedimentary Belt has only recently been noted (*see* Easton, this volume), and the authors assert that the recognition of other plutons, and associated mineralization, of this age is likely.

## MAGMATIC IRON-TITANIUM DEPOSITS

There are numerous iron and iron-titanium (deposit type 5) occurrences in the Central Metasedimentary Belt associated with the ca. 1245 million-year-old Lavant Suite and the ca. 1290 to 1270 million-year-old Nepheline Syenite Suite (*see* Figure 24.9b; Carter 1984). Mineralization consists principally of layers, lenses and pods of magnetite and titanite in the alkalic-intrusion-hosted deposits and magnetite, ilmenite and ilmenomagnetite in gabbro-hosted deposits (Carter 1984). Titanium abundances are generally highest in the gabbroic anorthosite-hosted deposits, locally ranging up to 15% $TiO_2$, with iron to titanium ratios ranging between 3:1 and 25:1 (Carter 1984).

### Syenite Association

Oxide mineralization occurs within syenite and nepheline syenite intrusions (e.g., Boulter, Fraser Lake, Bow Lake and Grattan deposits) and less commonly in syenite pegmatite dikes (e.g., Coltson Lake), all within the Bancroft Terrane. Mineralization generally consists of narrow parallel layers and lenses of disseminated to massive magnetite. The Grattan deposit (Carter et al. 1980) is the largest of these deposits, with a mineralized zone with a strike length of 2440 m varying from 4 to 10 m in width and a reserve of 9 million tons averaging 28% Fe and up to 1.8% $TiO_2$ (Carter et al. 1980; Carter 1984).

### Lavant Suite Association

Most of the iron-titanium oxide occurrences in the Central Metasedimentary Belt occur in Lavant Suite plutons (*see* Figure 24.9b). Mineralization consists of varied proportions of magnetite, ilmenite and magnetite containing exsolved ilmenite lamellae, generally as disseminated to massive lenses and layers parallel to the contacts of large intrusions. Mineralization is also conformable with foliation or rhythmic layering present in the host rocks, and occurs less commonly as randomly scattered lenses and pods of disseminated to massive oxides, or as diffuse zones of disseminated oxides (Carter 1984). The titanium content of the mineralized zones is extremely varied, ranging from 1 to 15% (Carter 1984).

The Newboro prospect is the largest iron-titanium deposit in southeastern Ontario (Carter 1984) and consists of 2 zones of well-layered, disseminated to massive ilmenite and ilmenomagnetite contained within an apparently unmetamorphosed anorthositic gabbro (Carter 1984). The zone is estimated to contain approximately

45 million tons of ore averaging 26.2% Fe and 6.6% $TiO_2$ (Carter 1984).

### Discussion

The intimate association of the iron-titanium deposit types with plutonic rocks, the association with high temperature mineral phases, and the preservation of igneous textures are consistent with a magmatic origin for the deposits (Gross 1967; Carter 1984). At some deposits (e.g., Newboro deposit), the iron-titanium oxides separated from the silicate portion of the gabbro-anorthositic magmas to form a late crystallizing fluid that was later injected into the upper levels of the intrusion when those phases were rigid enough to fracture (Krogh 1961; Gross 1967). A similar mechanism may account for the concentration of iron-titanium oxides in the syenites (Carter 1984).

# Metamorphism and Mineralization

## SKARN DEPOSITS

### Contact Metasomatic Iron (Type 6) Deposits

Contact metasomatic iron (type 6) deposits are widespread and are generally associated with ca. 1245 million-year-old Lavant Suite and Methuen Suite intrusions (*see* Figure 24.9c; Carter 1984). Although iron ore was produced from many of these deposits on a small scale, the Marmoraton deposit near the town of Marmora (*see* Figure 24.9c), and the Hilton deposit in Quebec, are the most significant deposits (Carter 1984). All occurrences contain less than 1% $TiO_2$ and are not enriched in base or precious metals (Carter 1984).

Mineralization consists of tabular, concordant bodies of disseminated to massive magnetite contained in irregular skarn zones. The skarn zones are developed within carbonate-rich metasedimentary rocks at or near the margins of intrusions (Carter 1984; Carter and Colvine 1985). The skarn mineralogy is varied, but generally consists of calcite, dolomite, diopside, hornblende, tremolite-actinolite, epidote, garnet, chlorite, talc, muscovite, biotite, apatite, pyrite, pyrrhotite and chalcopyrite (Carter 1984).

The Marmoraton deposit (*see* Figure 24.3; Rose 1958; Park 1966; Gross 1967) operated between 1955 and 1978, producing 25 365 853 tonnes of iron ore with an average grade of 42.8% Fe (Easton et al. 1986). The deposit consists of tabular, concordant lenses, layers and irregular zones of disseminated to massive magnetite in a heterogeneous skarn rock developed in a sequence of marble and subordinate siliceous metaclastic rocks at the margin of a syenite intrusion. Crude layering in the skarn and mineralized zones is oriented parallel to bedding in the metasedimentary rocks, and the skarn is inferred to be largely developed in the bedded rocks (Carter 1984).

Deposits of this type have been classified as pyrometasomatic, skarn, replacement, and contact metasomatic (Carter 1984). Although the deposits have been subsequently affected by regional metamorphism, with resultant recrystalization and deformation during the Elzevirian and Ottawan orogenies, the deposits are inferred to be genetically related to metamorphic and metasomatic effects at the margins of intrusions which cut carbonate metasedimentary rocks (Rose 1958; Giblin 1960; Park 1965, 1966; Carter 1984). The contact metasomatic origin is supported by: 1) the common spatial association with major intrusions; and 2) development within the surrounding marbles of unusual high temperature mineral assemblages typical of skarns.

### Skarn-Related Industrial Minerals

A variety of industrial minerals are related to contact metamorphism, and also are generally associated with the Lavant Suite and Methuen Suite plutons. Favourable host rocks are a necessary prerequisite for significant mineralization to occur. In the Madoc area, hydrothermal fluids associated with the high-fluorine Deloro Granite have formed significant wollastonite, talc and dolomite deposits, including the Canada Talc Mine which has been in operation since 1893. It is not known whether the composition of the Deloro Granite played an important role in the formation of these deposits. In other parts of the Central Metasedimentary Belt, higher metamorphic grade may prevent the discrimination of contact versus regional metamorphic deposits.

## REGIONAL METAMORPHISM

Regional metamorphism appears to be a far more significant factor affecting the formation of industrial mineral prospects than metallic prospects. In metallic deposits, regional metamorphism may be significant in improving the milling properties of the ore, or in producing high-grade zones, however, in industrial mineral deposits, in many instances it forms the ore. For example, regional metamorphism of suitable aluminous horizons within the Flinton Group was responsible for the formation of sillimanite, kyanite, feldspar and mica prospects within this unit in the Clare River and Ardoch–Ompah areas. Optimum host rock types, metamorphic fluid composition, and the intensity and length of metamorphism were the critical factors in the formation of these industrial mineral prospects. Other examples include the formation of talc bodies in metamorphosed ultramafic and metagabbroic rocks in the Grimsthorpe Domain. In this instance, contact metamorphism associated with the emplacement of tonalite plutons in the area may have been an important preconditioning in the development of these deposits.

# Controls on Gold Mineralization

Epigenetic gold-bearing quartz ± carbonate veins (type 7 deposits) occur throughout the lower metamorphic grade parts of the Central Metasedimentary Belt, predominantly within the Elzevir, Mazinaw and Sharbot Lake terranes (*see* Figure 24.9d). In most of the deposits, gold is associated with subordinate arsenic, antimony, copper, lead and zinc mineralization (Carter and Colvine 1985). Silver is usually present in only minor amounts, and gold to silver ratios are consistently greater than 1 (Carter 1984). The gold deposits occur in clusters, probably comparable to the camps within the Superior Province, each characterized by slightly different attributes (Table 24.6). This camp nomenclature will be used henceforth. All gold-quartz veins are structurally controlled, although in several camps there is a spatial association with specific rock types.

Several factors seem to control the distribution of gold-bearing veins within the Central Metasedimentary Belt. These include: 1) many veins occur near the greenschist-amphibolite facies transition, generally hosted in upper greenschist facies rocks; 2) most prospects are hosted in either mafic metavolcanic or metagabbroic rocks, or in dolomitic marbles; 3) many prospects are spatially associated with regional deformation zones; many of these deformation zones are associated with terrane and domain boundaries; 4) trondhjemite dikes, commonly cupriferous, are associated with some of the gold prospects; 5) gold is commonly fine grained and contained within sulphides and along grain boundaries; visible gold is rare within the vein systems; and 6) in the Mazinaw Terrane, mineralized veins are commonly located near the unconformity between the Mazinaw Group and the ca. 1150 million-year-old Flinton Group. It is still unclear if all of these veins are related to a single gold mineralization event, or if several ages of gold mineralization are present. The significance of these factors, with examples, are outlined briefly below, followed by a brief description of the various gold camps. LeBaron (1991) summarized recent exploration for gold in southeastern Ontario. Carter (1984), Papertzian (1984), Malzack et al. (1985) and LeBaron (1991) describe the more significant prospects.

## METAMORPHISM

Figure 24.9d outlines the distribution of gold occurrences within the Central Metasedimentary Belt, as well as the greenschist-amphibolite facies boundary. All of the past-producers and all significant prospects are found near this metamorphic transition. In many areas, this boundary is structurally controlled (e.g., the Robertson Lake mylonite zone), and in such instances, mineralized veins are generally hosted in rocks that have been subjected to retrograde greenschist facies metamorphic conditions (e.g., Lavant–Darling camp, Addington–O'Donnell camp). The time of vein development and peak metamorphism is not known in most cases. In the Mazinaw Terrane, 2 regional metamorphic episodes are documented. Mineralized quartz-tourmaline veins are younger than the ca. 1230 to 1180 million-year-old metamorphism associated with the Elzevirian Orogeny, however, it is unclear if the veins are late-, syn- or post- the second metamorphic event at ca. 1150 to 1090 Ma. Some gold occurrences associated with the Swamp Lake shear zone in the Mazinaw Terrane (e.g., Boerth Mine) occur in middle amphibolite facies rocks, however, this metamorphic assemblage reflects the second metamorphic event, and it is possible that initial mineralization occurred at lower grade. In the Sharbot Lake Terrane, the veins are post–peak Elzevirian metamorphism, however, the time of vein emplacement is not well constrained, and occurred sometime between 1180 and 980 Ma.

## VEIN CHARACTERISTICS

Most gold deposits are present in quartz or carbonate veins containing a variety of accessory minerals, the most common of which is iron carbonate. Pyrite is present in all gold vein systems, with arsenopyrite being more abundant in veins cutting metavolcanic and metagabbroic rocks. Overall, arsenic seems to be more abundant in the Mazinaw Terrane, as shown on regional geochemical maps (*see* Easton, this volume). Antimony sulphides are most common in veins cutting carbonates (Carter 1984). In the Mazinaw Terrane, tourmaline and actinolite are commonly associated with gold-bearing quartz veins, and rare scheelite is present. Plagioclase and orthoclase occur as accessories in the Cordova camp.

Two main types of gold-bearing vein systems are recognized: 1) strata-bound, concordant to discordant quartz-dolomite vein arrays in dolomitic marble (Figure 24.17); and 2) concordant to discordant quartz and quartz-ankerite vein arrays (Figures 24.18 and 24.19; Carter 1984). Deposits of the first type are generally small, with no history of production, and consist of strata-bound networks of narrow (less than 10 cm) quartz, quartz-dolomite, and dolomite veins, pods, and lenses which cut dolomitic marbles. Mineralization consists of erratically disseminated chalcopyrite, pyrite, tetrahedrite, bornite, boulangerite, jamesonite, sphalerite, galena and rare arsenopyrite (Carter 1984; Carter and Colvine 1985).

Quartz and quartz-ankerite vein arrays discordant to foliation or layering in the host rocks are present in many of the concordant vein arrays and appear to be syn- to late-deformational (Carter 1984). Gold production has been attained from several deposits of this type, principally the Cordova and Deloro camps. There can be a gradation from concordant dolomite-rich, to quartz-dolomite lenses, to crosscutting quartz veins; however, crosscutting veins only rarely form the principal vein structure (e.g., Boerth, Marble Lake camp, Carter 1984). Veins having discordant geometry occur within several structural settings (Carter 1984): 1) in shear zones cutting gabbro (Cordova camp), granite (Malone camp) or mafic metavolcanic rocks; 2) in fractures crosscutting marble and siliceous clastic metasedimentary rocks; and 3) concordant to discordant lenses in shear zones along the basal unconformity of the Flinton Group (Marble Lake camp).

The spatial association of gold deposits with metavolcanic rocks has been cited in support of the derivation of metals from the metavolcanic rocks or from strata-bound sulphide deposits occurring at the top of the volcanic successions (Carter 1984). However, many gold deposits are hosted in carbonate rocks or volcaniclastic metasedimentary rocks, and a direct link with volcanism as a source of metals cannot be established. Alternatively, hydrothermal fluids may have been focussed into these rock successions because of their favourable mechanical and chemical characteristics.

## REGIONAL STRUCTURAL CONTROL

Gold deposits in the Central Metasedimentary Belt occur within, or are spatially associated with, deformation zones,

**Table 24.6.** General characteristics of gold deposits in the Central Metasedimentary Belt, Grenville Province (*compiled from* Carter 1984).

| Attribute | Deposit Name | | | | | |
|---|---|---|---|---|---|---|
| | **Malone Camp** | **Cordova Camp** | **Bannockburn Camp** | **Addington–O'Donnell Camp** | **Lavant–Darling Camp** | **Marble Lake–Harlowe Camp** |
| Example | Deloro, Sovereign, Cook, Dean Williams | Cordova, Ledyard, Pearce | Craig deposit, Mono gold, St. Joe, Bannockburn gold, Sophia mine | Addington mine, O'Donnell occurrences | Lavant–Darling deposits | Boerth, Gough, Star, Ore Chimney, Helena mines and deposits |
| Vein Mineralogy | q, asp | q, ank, plag, or, py, (po), (Au) | q, py, cp, (po) | q, py, (asp), (po), (cp), (gn), (sp), (Au), (shee) | q, q-dol, dol veins, pods, lenses; (Cu-Sb-Au-Ag); cp, py, tn, bn, bl, jm, sp, gn, (asp) | q-tour, q-act, q-dol, dol veins, pods, lenses; py, (asp), (sp), (gn), (shee), (Au), (bm), (cp), (bl), (po), (tn) |
| Alteration | Fe-carb, ser, asp | Fe-carb, elevated potassium | Fe-carb, elevated potassium | asp, Fe-carb, (tour), (bio), (mag), (il), bio, gt | Silicified breccia zones, dol | tour; some sulphide mineralization in wall rock |
| Host Rock | Calcitic and dolomitic marble intruded by 1245 Ma Deloro Granite | 1242 Ma gabbro | Mafic metavolcanic and volcaniclastic metasedimentary rocks (ca. 1250 Ma) within contact aureole of Lingham Lake diorite-gabbro intrusion | Sheared mafic metavolcanic rocks; may or may not be spatially associated with rocks of Flinton Group | Sheared mafic metavolcanic rocks, dolomitic and calcitic marbles; sheared Addington Granite (1241 Ma) and Lavant Gabbro | Flinton Group; sheared conglomerate, black shale<br>Mazinaw Group: dolomitic marble, sheared mafic to intermediate metavolcanic rocks |
| Metamorphic Grade | Near greenschist-amphibolite transition | Near greenschist-amphibolite transition | Hosted in greenschist facies rocks; near greenschist-amphibolite transition | Hosted in greenschist facies rocks; near greenschist-amphibolite transition | Hosted in greenschist facies rocks; near greenschist-amphibolite | Near greenschist-amphibolite transition; pre–Flinton Group |
| Spatially Associated Structure | North- to northeast-striking deformation zone | East- to northeast-striking deformation zone | North-northwest-striking deformation zone coincident with domain boundary | North-striking deformation zone coincident with terrane boundary | North-striking deformation coincident with terrane boundary | Northeast-striking thrust faults and reactivated thrust faults and post–Flinton Group shear zones and fractures |
| Age Constraint on Host Structure | <1245 Ma<br>Gold younger than 1245 Ma | <1245 Ma<br>Gold younger than 1240 Ma | 1270 to 1230 Ma possible early development of Grimsthorpe domain boundary shear zone<br>Gold younger than 1270 Ma | 1270 to 1230 Ma possible early development of Elzevir–Mazinaw terrane boundary<br>Terrane boundary cut by 1090 Ma syenite dike<br>Gold younger than 1270 Ma and older than 1050 Ma | Gold younger than 1240 Ma | Early (<1250 Ma) shearing, thrusting and metamorphism of Mazinaw Group<br>Post 1160 Ma to ca. 1050 Ma shearing and metamorphism of Flinton Group and underlying Mazinaw Group<br>Gold younger than 1250 Ma, possibly younger than 1150 Ma |

Table 24.6. General characteristics of gold deposits in the Central Metasedimentary Belt, Grenville Province (*compiled from* Carter 1984).

| | Deposit Name | | | | | |
|---|---|---|---|---|---|---|
| **Attribute** | **Malone Camp** | **Cordova Camp** | **Bannockburn Camp** | **Addington–O'Donnell Camp** | **Lavant–Darling Camp** | **Marble Lake–Harlowe Camp** |
| Host Terrane | Elzevir | Elzevir | Elzevir | Mazinaw | Sharbot Lake | Mazinaw |
| Tectonic Setting | Spatially associated with fault within Belmont Domain | Spatially associated with fault within Belmont Domain | Within Grimsthorpe Domain along domain boundary with Belmont Domain | Along boundary between Mazinaw and Elzevir terranes | Along boundary between Sharbot Lake and Mazinaw terranes | Spatially associated with<br>a) basal Flinton Group unconformity; b) early thrust faults at base of Flinton Group;<br>c) unconformity and thrust faults that were reactivated during post–Flinton deformation and metamorphism;<br>d) late trondhjemite dikes |

Abbreviations: act - actinolite; Ag - silver; ank - ankerite; asp - arsenopyrite; Au - gold; bio - biotite; bl - boulangerite; bm - bismuthinite; bn - bornite; cp - chalcopyrite; Cu - copper; dol - dolomite; Fe-carb - iron-rich carbonate; gn - galena; gt - garnet; il - ilmenite; jm - jamesonite; mag - magnetite; or - orthoclase; plag - plagioclase; po - pyrrhotite; py - pyrite; q - quartz; Sb - antimony; ser - sericite; shee - scheelite; sp - sphalerite; tn - tetrahedrite; tour - tourmaline

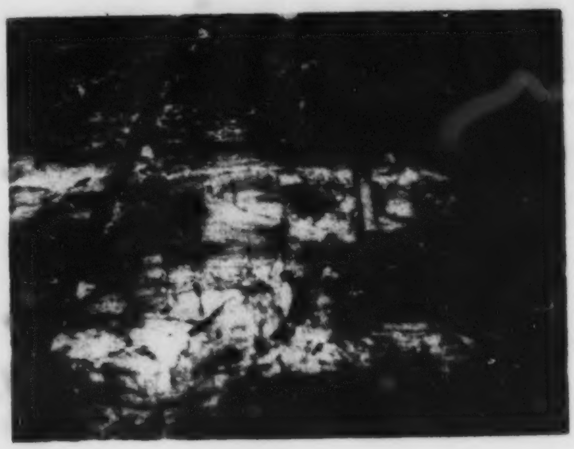


**Figure 24.17.** Concordant to discordant quartz vein array in dolomitic marble of the Mazinaw Group (recessive) at the Dome gold occurrence, Mazinaw Terrane. Hammer handle is 30 cm long.

as follows: 1) shear zones within domains (e.g., Cordova and Malone camps, Elzevir Terrane; Marble Lake camp, Mazinaw Terrane; 2) shear zones which define domain boundaries (e.g., Bannockburn camp, Belmont–Grimsthorpe domain boundary); and 3) shear zones which define terrane boundaries (e.g., Addington–O'Donnell camp, Elzevir–Mazinaw terrane boundary; Lavant–Darling camp, Mazinaw–Sharbot Lake terrane boundary). Deposits in the Marble Lake camp, Mazinaw Terrane, are spatially associated with shear zones near or below the basal unconformity of the Flinton Group; these shear zones may have served to localize deposition of the Flinton Group (Easton and Ford 1991).

The shear zones provided dilatant zones favourable for localization of veins, and the occurrence of many of the shear zones along contacts between different rock types suggests that ductility contrasts were important in controlling shear zone and vein localization (Carter 1984). Veins in carbonate rocks are less well developed, possibly because of the more ductile character of calcite-rich rocks (Carter 1984).

## CAMP-SCALE ALTERATION PATTERNS

Alteration assemblages generally reflect the bulk composition of the host rock. Mafic metavolcanic or metagabbroic rocks are replaced by iron-bearing carbonate and silica. Dolomitic and calcitic host rocks may be replaced by iron-bearing carbonate minerals. Granitic host rocks are sericitized. In the Marble Lake area, alteration assemblages adjacent to gold-quartz veins are superimposed upon a pre-existing (Elzevirian) metamorphic assemblage (Easton and Ford 1991). The relation to alteration and regional metamorphism has not been established for the other camps.

## ASSOCIATION WITH TRONDHJEMITE DIKES

In several camps, there is a spatial association with late trondhjemite dikes and small plutons. In the Mazinaw

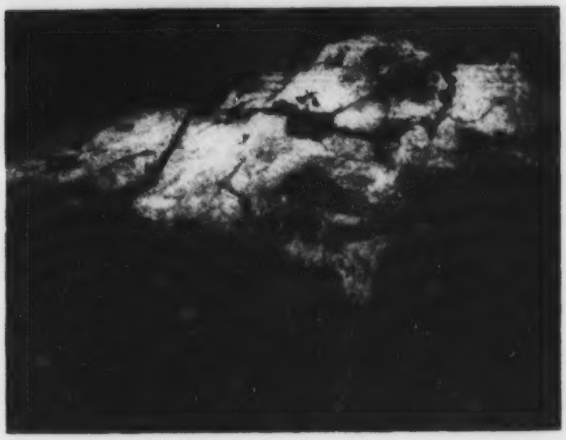


**Figure 24.18.** Discordant gold-bearing quartz vein crosscutting siliclastic, volcanic-derived, metasedimentary rocks at the Mono gold prospect, Bannockburn camp, Elzevir Terrane. Hammer handle is 30 cm long.

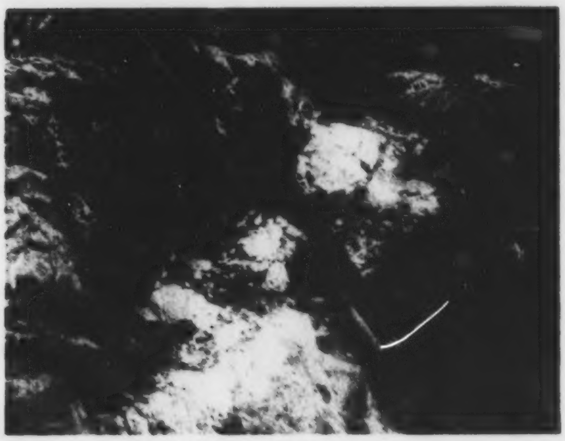


**Figure 24.19.** Discordant gold-arsenopyrite-pyrite vein cutting gabbroic rocks along the western margin of the Deloro Granite, south of the Deloro gold mine, along Highway 7. Hammer handle is 30 cm long.

Terrane in particular, copper-gold deposits such as the Helena Mine are intimately associated with these dikes, even though the veins are most abundant in the dolomitic marbles that the dikes intrude. In the Mazinaw Terrane, the dikes are cupriferous (300 to 4400 ppm, Papertzian 1984; Easton 1992). Easton (1992) noted that these dikes have similarities to porphyry copper-gold systems found in the Canadian Cordillera and suggested an exploration model for the Mazinaw Terrane based on these similarities. The age of these dikes is not well constrained. In the Mazinaw Terrane, they could be as old as 1229±11 Ma (age of "dacite dikes" near Elzevir Tonalite, Connelly et al. 1987), or as young as 1160 Ma (pre–Flinton Group deposition). Trondhjemite dikes are also present in the Bannockburn (Easton, unpublished data) and Lavant–Darling camps (Easton 1988a).

## AGE OF GOLD MINERALIZATION

Constraints on the maximum age(s) of gold deposit formation are provided by crosscutting relationships between mineralized shear zones and host intrusions. The maximum age of the Cordova and Malone camps is 1240 Ma. Gold deposits in the Addington–O'Donnell camp are younger than 1270 Ma, and most likely younger than 1240 Ma. In the Mazinaw Terrane, gold-copper-bearing quartz-tourmaline veins cut the Flinton Group, implying a maximum age of ca. 1155 Ma. However, these crosscutting veins may represent remobilization of mineralization during the younger regional metamorphism (Ottawan Orogeny), and the main mineralization event may have occurred post–ca. 1180 Elzevirian metamorphism.

If alteration assemblages in the Addington–O'Donnell camp truly retrograde peak metamorphic assemblages along the terrane boundary shear zone, then gold mineralization in this area could be as young as 1050 Ma, which is the $^{40}Ar/^{39}Ar$ blocking temperature of hornblende in this area (*see* Easton, this volume). Similarly, if the Robertson Lake mylonite zone developed at the same time as late extensional faults in the Central Gneiss Belt and the Bancroft Terrane, then gold mineralization in the Lavant–Darling camp could be as young as 1040 Ma (*see* Easton, this volume), assuming that emplacement of the veins and retrograde metamorphism were related.

If one assumes that a single gold mineralizing event occurred, it most likely occurred at ca. 1180 to 1160 Ma. This would make mineralization post–peak Elzevirian metamorphism, which is coincident with regional thrusting and shearing throughout the Grenville Province in Ontario. In the Mazinaw Terrane, it is consistent with the observation that an early gold mineralization event (particularly in dolomite-marble-hosted concordant vein arrays, e.g., Dome and Payrock occurrences, *see* Figure 24.17) occurred prior to Flinton Group deposition. The high metamorphic grade of the Flinton Group metamorphism would be sufficient to cause local remobilization to form crosscutting veins; perhaps the high boron content of Flinton Group sediments (Easton, unpublished data) assisted this process.

## ELZEVIR TERRANE OCCURRENCES

### Cordova Area

Gold deposits in the Cordova camp occur as quartz-feldspar-ankerite veins within shear zones in the northwestern margin of the ca. 1242±3 million-year-old Cordova Gabbro (Thomas and Cherry 1981; Springer 1983; Carter 1984; Easton et al. 1986; Davis and Bartlett 1988). Mineralization consists of sparsely disseminated pyrite and minor pyrrhotite; gold occurs in the pyrite (Carter 1984). Attitudes of the shear zones in the gabbro are varied; however, the main mineralized zones at both the Cordova and Ledyard deposits strike east and dip moderately to steeply to the south (Carter 1984). Within the shear zones, the gabbro is strongly foliated and altered to biotite and chlorite (Carter 1984).

### Malone Area

In the Malone area, 2 types of gold mineralization are present. The first is gold-arsenic mineralization, which includes the past-producing Deloro, Sovereign, Cook and Dean Williams deposits, and is confined to quartz veins within shear zones or fractures in the heterogeneous zone of gabbro-diorite-syenite-granite that occurs along the western margin of the Deloro Granite (*see* Figure 24.19), adjacent to the contact with a thick succession of calcite and dolomite marbles. The shear zones hosting the gold mineralization vary from a few centimetres to 3 m thick, generally strike north, parallel to the pluton margin, and dip moderately to the west (Carter 1984). All the rocks in the contact zone are extensively altered, especially near the gold-arsenic deposits. Schistose wall rock is replaced by iron carbonate, arsenopyrite and masses of diopside and mica and irregular lenses and discontinuous veins of white to glassy quartz and coarse-grained iron carbonate cut the sheared wall rock (Carter 1984).

The second type of gold mineralization, and more significant economically, is gold disseminated in sheared and sericitized granitic rocks within, or neighbouring, the 1241±2 million-year-old Deloro Granite (van Breemen and Davidson 1988b). The Dingman prospect is the best known deposit of this type (LeBaron 1991). Other examples are the Gatling Five Acre Mine and the Ackerman prospect. In the Dingman prospect, mineralization is hosted in sheared, sericitized and locally hematitized granite of the Malone pluton. Pyrite, chalcopyrite and magnetite are spatially associated with the gold. Mineralization occurs both within the granite and in crosscutting quartz stringers. The prospect lies near the intersection of 2 regional fault trends, but is not located at a domain or terrane boundary.

### Bannockburn Area

Several prospects are associated with the regionally northwest-striking carbonate-metavolcanic contact in the Bannockburn area. Most prospects have been re-examined within the last decade (LeBaron 1991), and include the Bannockburn gold prospect, the Mono prospect (*see* Figure 24.18), the St. Joe occurrence and the Sophia Mine. The carbonate-metavolcanic contact marks the Belmont–Grimsthorpe domain boundary, and may represent an early (ca. 1230 to 1180 million-year-old) thrust zone. In addition, several gold prospects occur within deformed mafic metavolcanic rocks east of the boundary; these include the Cooper, Dillman and Craig prospects and the Gilmour Mine. LeBaron (1991) and Malzack et al. (1985) described many of these prospects, but their regional geologic setting is not well understood.

## MAZINAW TERRANE OCCURRENCES

### Marble Lake–Harlowe Area

The structural setting of deposits in the Marble Lake–Harlowe camp is complex (*see* Easton, this volume). Deposits occur within or in proximity to: 1) shear zones that developed during a ca. 1180 to 1160 million-year-old deformation and metamorphic event which affected metavolcanic and metasedimentary rocks of the Mazinaw Group; and 2) the unconformity at the base of the Flinton Group which developed at roughly 1155 Ma. Some quartz vein arrays are concordant to pre–Flinton Group fabrics within the shear zones, others, particularly quartz-tourmaline veins, are discordant and cut both Mazinaw and Flinton group strata. Relationships are further complicated by the likelihood that shear zones along, or near, the basal unconformity locally may have been reactivated during deposition of the Flinton Group and the metamorphic event that affected the Flinton Group (*see* Easton, this volume; Easton and Ford 1991). In addition, there is an association of some of the older Mazinaw Group dolomite-hosted vein arrays with cupriferous (300 to 4400 ppm) trondhjemite dikes and small plutons (Easton 1992). Descriptions of many of these deposits are given in Meen (1944), Papertzian (1984), Carter (1984), Malzack et al. (1985) and LeBaron (1991).

### Addington–O'Donnell Area

The Addington and O'Donnell deposits occur within sheared mafic metavolcanic rocks along the deformation zone that marks the boundary between the Elzevir and Mazinaw terranes. Metamorphosed quartz arenite and metaconglomerate of the Flinton Group are structurally juxtaposed against the metavolcanic rocks at the Addington deposit. Gold mineralization occurs within shear zones parallel to the terrane boundary, which cut both metavolcanic and metasedimentary rocks (Carter 1984). Within the shear zones, the metavolcanic rocks are altered to biotite and garnet and mineralization consists of localized quartz stringers (Carter 1984). The veins occur parallel to, and at an angle to, the shear foliation (Carter 1984). Pyrite, accompanied by minor arsenopyrite, pyrrhotite, chalcopyrite, sphalerite and galena are the principal sulphide minerals (Carter 1984). Native gold occurs in fractured arsenopyrite and is spatially associated with magnetite (Bell 1949).

## SHARBOT LAKE TERRANE OCCURRENCES

Gold occurrences in the Sharbot Lake Terrane consist of strata-bound, copper-, antimony- and gold-bearing veins located within the Robertson Lake mylonite zone, particularly where it transects the Lavant Gabbro (Carter 1984)..Rocks within the shear zone have been mylonitized and exhibit retrograde greenschist facies assemblages (Carter 1981; Easton 1988a). Mineralization consists of erratically dispersed grains and blebs of tetrahedrite, chalcopyrite, pyrite and bornite contained in a complicated network of quartz-dolomite veins, pods and lenses (Carter 1984), and occurs in both veins of dolomite and quartz. The quartz veins cut all other features in the marbles, have sharply defined contacts, and are, themselves, relatively undeformed. Carter (1984) distinguished 2 sulphide mineral assemblages: 1) disseminated euhedral pyrite; and 2) anhedral chalcopyrite, tetrahedrite, bornite and occasional pyrite. The first assemblage is spatially associated with the early dolomite layers and lenses, and is nearly devoid of quartz veins. The chalcopyrite-tetrahedrite-bornite±pyrite assemblage occurs either as dispersed intergranular grains within quartz and quartz-dolomite veins and dolomite layers and lenses, or as disseminated grains within healed fractures in quartz or quartz-dolomite veins and dolomite layers, lenses and veins (Carter 1984). Gold is associated with both pyrite and tetrahedrite (Carter 1981).

## Mineralization Related to Post–Grenville Events

### URANIUM-, MOLYBDENUM- AND RARE ELEMENT-BEARING PEGMATITES, SKARNS AND VEINS (TYPE 8 DEPOSITS)

All significant uranium-, molybdenum- and rare element-bearing pegmatites, skarns, and fluorite-apatite-calcite veins are confined to areas of middle to upper amphibolite facies metamorphism, predominantly within the Bancroft Terrane (*see* Figure 24.9e). These deposits are all spatially and temporally associated with each other (Figure 24.20), and are interpreted to be cogenetic (Lentz 1991). Historically, apatite, molybdenum, phlogopite, muscovite, feldspar, fluorite and uranium have been mined from these deposits. Thorium, yttrium, beryllium, rare earth elements and niobium also are concentrated in the deposits (Ellsworth 1932; Lang 1952; Shaw 1958; Rose 1960; Hewitt 1967a, 1967b). The deposits occur in several rock types and habits, including (Table 24.7; Satterly 1957; Masson and Gordon 1981): 1) zoned and unzoned granitic pegmatites; 2) sulphide-rich siliceous marbles; 3) calc-silicate skarns developed in marble and pyroxenite; and 4) calcite- and pyroxene-bearing veins.

Examples of radioactive granitic pegmatites include the Faraday (Madawaska), Bicroft, Greyhawk, Dyno and Halo deposits, all of which produced uranium (Robertson 1981). The pegmatites are subdivided into zoned and unzoned granite pegmatites of both mineralogically simple and complex types (*see* Table 24.7; Satterly 1957; Hewitt 1967a, 1967b). Pegmatites are either oxidized, hematite bearing and red coloured, or less oxidized and white (Lentz 1991). White pegmatites in the Mount Laurier and Renfrew areas occur within marble-dominated successions (Tremblay 1974; Charbonneau and Jonasson 1976). Red pegmatites in the Renfrew area occur in amphibolite or gneiss (Charbonneau and Jonasson 1976). Pegmatites crystallized between 600°C and 630°C at approximately 400 MPa (Lentz and Kretz 1989).

Complex pegmatites usually contain uranium-, molybdenum- and rare earth element-bearing minerals and consist of quartz, microcline perthite, plagioclase, and may contain albite, biotite, calcium-rich pyroxene, hornblende, magnetite, pyrite, calcite, fluorite and numerous rare accessory minerals (Masson and Gordon 1981; Carter 1984;

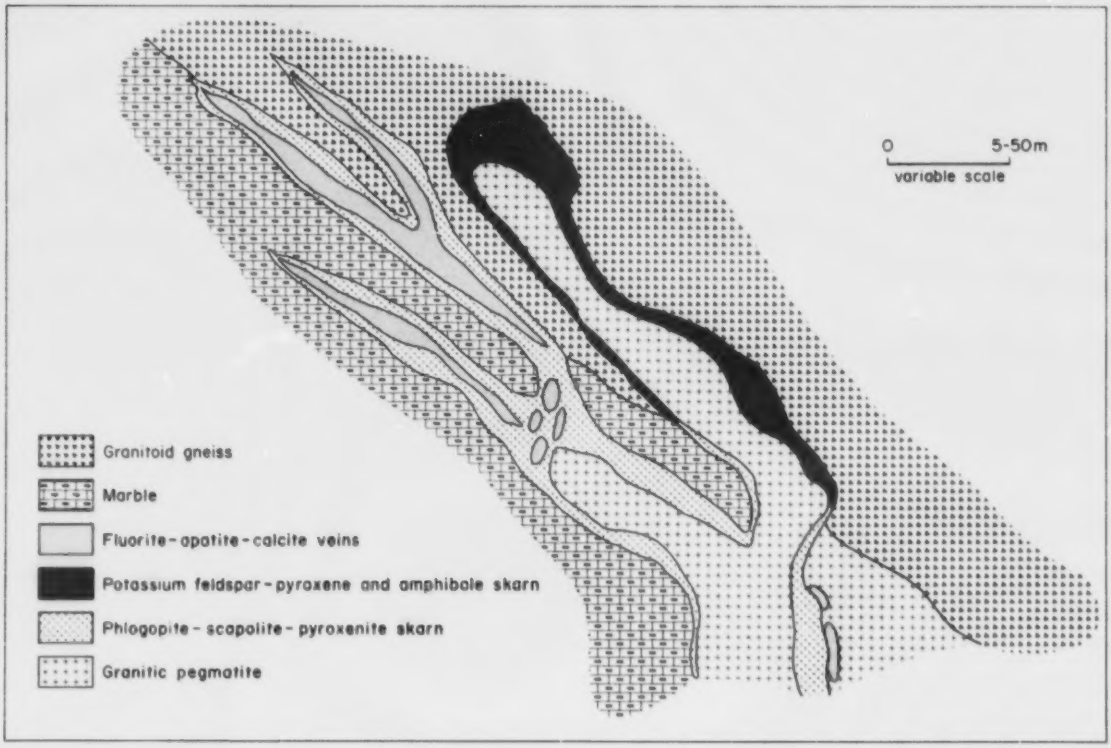

0
5-50m
variable scale
Granitoid gneiss
Marble
Fluorite-apatite-calcite veins
Potassium feldspar-pyroxene and amphibole skarn
Phlogopite-scapolite-pyroxenite skarn
Granitic pegmatite

**Figure 24.20.** Schematic diagram depicting the spatial relationship between uranium, molybdenum and rare earth element-bearing granite pegmatites, skarns and veins in the Grenville Province (*modified from* Lentz 1991).

**Table 24.7.** Types of uranium-bearing deposits in the Central Metasedimentary Belt, Grenville Province (*modified from* Satterly 1957; Lentz 1991).

**Granite Pegmatites**

- A. Zoned pegmatite
    - 1. Simple mineralogy
    - 2. Complex mineralogy
- B. Unzoned pegmatite
    - 1. Simple mineralogy
    - 2. Complex mineralogy

**Metasomatic Skarn**

- A. Marble-hosted
    - 1. Dolomitic marble
    - 2. Calcitic marble
- B. Metapyroxenite-hosted

**Hydrothermal Fluorite-Apatite-Calcite Veins**

- A. Gneiss-hosted
- B. Amphibolite-hosted
- C. Marble-hosted
- D. Metapyroxenite-hosted

Lentz 1991). Common rare minerals include uraninite, uranothorite, euxenite, fergusonite, cyrtolite, pyrochlore, samarskite, allanite, andradite and titanite (Lentz 1991). Simple pegmatites consist of quartz, microcline perthite, plagioclase and generally do not contain mineralization (Lentz 1991).

Both types of pegmatites occur as long, narrow dikes or sills up to 1 km long and up to 50 m wide (Masson and Gordon 1981). In the Bancroft Terrane, most of the uraniferous pegmatites occur within regional deformation zones, such as the Central Metasedimentary Belt Boundary Zone (*see* Easton, this volume), in northeast-trending deformation zones (Bright 1977), or near the flanks of large gneiss domes (e.g., Harvey–Cardiff Arch; *see* Easton, this volume) and plutons (Masson and Gordon 1981).

Uranium mineralization consists principally of disseminated uraninite and uranothorite, with minor to locally abundant allanite, thorite, thorianite, secondary uranophane, and columbates of titanium and tantalum (Ellsworth 1922; Freeman 1936; Thomson 1943; Satterly 1957; Robinson and Hewitt 1958; Hewitt 1967b; Steacy et al. 1973; Bright 1977, 1979; Gordon et al. 1981; Masson and Gordon 1981). Accessory minerals include magnetite, zircon, titanite, apatite, tourmaline, hematite, calcite, fluorite, anhydrite, pyrite, marcasite, pyrrhotite and molybdenite (Carter 1984). Molybdenite occurs randomly distributed as coarse flakes within the pegmatite in fractures, veins and stringers (e.g., Kellar and Liedtke prospects; Karvinen 1973). The mineralization can occur in sheared, fractured parts of the pegmatites, usually in altered zones near the contacts with the country rocks (Masson and Gordon 1981). Mineralized zones can occur randomly within the pegmatites, but in the larger pegmatites, the mineral occurrences form lenticular zones along, and oriented parallel to, the contacts of the pegmatite, or are medial (Carter 1984). Highest uranium grades are typically indicated by increased amounts of hematite, magnetite and mafic wall rock xenoliths (Carter 1984). Associated alteration includes silicification, hematitization and albitization (Carter 1984).

Radioactive uranium- and molybdenum-bearing skarns are developed in marbles, clinopyroxenite and amphibolite and occur as marginal zones to granitic pegmatites, as veins, or as pervasive replacement zones (*see* Table 24.7; Lentz 1991). The skarn zones are coarse grained and are radioactive, whereas the host marbles or clinopyroxenites are not radioactive (Lentz 1991).

Marble-hosted skarn assemblages can develop as conformable zones along contacts between marble and paragneiss (Figure 24.21; Karvinen 1973; Carter 1984). Marbles in contact with mineralized skarns are calcitic, with minor amounts of dolomite, muscovite, quartz, biotite, phlogopite, albite and sometimes diopside-albite-apatite-titanite assemblages (Vokes 1963; Karvinen 1973). Dolomitic and calcitic marbles are locally altered to silica-rich assemblages ("silicated marbles") of diopside, calcite, tremolite, phlogopite and sulphides (Lentz 1991). Skarn assemblages formed between 525°C and 575°C and at 400 MPa, with $X(CO_2)$ between 0.1 and 0.3 (Lentz 1991). The marble-hosted skarn assemblages probably formed by metasomatic introduction of silica, alkalis, alumina and volatiles (Lentz 1991).

Clinopyroxenite-hosted skarns are replacement zones and veins consisting of coarse-grained calcium-rich pyroxene, plagioclase, potassium feldspar, scapolite, phlogopite and minor calcite (*see* Figure 24.21 and Table 24.7). The skarns occur as strata-bound lenses within sequences of granite, quartzite, or biotite and hornblende gneisses, interlayered with marbles (Karvinen 1973; Carter 1984). Marble is not common, or is absent, in proximity to the pyroxene-albite skarn assemblages (Carter 1984). Pegmatite dikes and sills are present at some deposits. In others, such as the Hunt Mine and Matthews–McMahon deposit, the skarn zones occur at the contacts between pegmatite and marble or paragneiss (Karvinen 1973; Carter 1984). Coarse-grained flakes of molybdenite are erratically dispersed in the dark green, massive, coarse-grained skarn assemblage, commonly associated with pyrite, pyrrhotite and minor uranium (Karvinen 1973; Carter 1984). Average molybdenite contents of the deposits range up to 1.5% $MoS_2$, but quantification is difficult because of erratic distribution of the molybdenite (Karvinen 1973; Carter 1984). Accessory pyrite and pyrrhotite locally form massive lenses within, and gradational into, the skarn assemblages (Karvinen 1973). In scapolite-rich skarns, molybdenite is preferentially concentrated in the sulphide-rich lenses or occurs as sparsely scattered flakes in sulphide-poor parts of the skarn (Karvinen 1973). In feldspar-rich skarns, molybdenite is more abundant in the diopside-rich parts of the skarn (Karvinen 1973).

Amphibolite-hosted skarns are texturally and mineralogically similar to the pyroxenite-hosted skarns and host coarse-grained molybdenite concentrated in pegmatitic

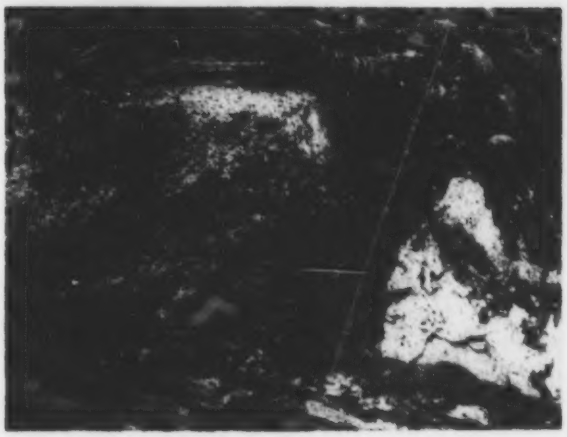


**Figure 24.21.** Clinopyroxenite skarn in calcitic marbles of the Denna Lake Structural Complex near Minden, Ontario. Hammer handle is 30 cm long.

parts of the amphibolites (Karvinen 1973; Carter 1984; Lentz 1991). Secondary assemblages consist of hedenbergite, scapolite, minor titanite, apatite, iron sulphides and rare quartz (Karvinen 1973). In scapolite-rich gneiss, scapolite is largely replaced by albite or oligoclase (Karvinen 1973). The amphibolites normally occur as conformable layers within mixed sequences of carbonate and siliceous clastic metasedimentary gneisses, usually along marble-paragneiss contacts (Karvinen 1973). The marbles consist predominantly of calcite, although minor amounts of dolomite, diopside, mica, graphite and quartz also occur (Karvinen 1973).

Fluorite-apatite-calcite veins occur in shear zones and along geological contacts. The veins are generally less than 1 m thick and range up to few hundred metres in strike length. Fluorite, apatite and calcite are the principal vein minerals; calcium-rich pyroxene, calcium- and sodium-rich amphibole, biotite, feldspar, titanite and radioactive minerals occur along vein selvages (Lentz 1991). Marble-hosted veins resemble the "silicated marble" skarn, but are fracture filling rather than replacement zones (e.g., Cardiff uranium-fluorite occurrence; Lentz 1991). Clinopyroxenite-hosted veins are also similar to skarn, but have less plagioclase, potassium feldspar and scapolite; fine- to coarse-grained calcium-rich pyroxene, tremolite and phlogopite occur on vein selvages, accompanied by minor molybdenite, pyrite and pyrrhotite (Lentz 1991). Selvages adjacent to gneiss-hosted veins contain potassium feldspar, calcium-rich pyroxene, calcium- and sodium-rich amphibole, biotite, pyrite and/or magnetite (e.g., Richardson uranium deposit; Lentz 1991). Hornblende and biotite, accompanied by minor scapolite, plagioclase or potassium feldspar, are principal selvage minerals adjacent to veins in amphibolite (Lentz 1991).

### Discussion

The pegmatites, skarns and veins are posttectonic, and formed between 1060 and 1020 Ma (Lentz 1991; Lumbers et al. 1990), some 30 million years after the Ottawan Orogeny (ca. 1100 Ma; Easton 1986). Deposits are concentrated in the Central Metasedimentary Belt Boundary Zone and the Bancroft Terrane which contain faults that extend deep into the crust (*see* Easton, this volume) and which may have served as fluid channelways. The genetic process may have involved high-grade regional metamorphism, production and emplacement of pegmatite melts, and regional extension (Masson and Gordon 1981; Carter 1984). A genetic relationship with intrusions has not been favoured (Carter 1984). Alternatively, these deposits may have formed from magmatic-hydrothermal fluids derived from crystallizing granitic pegmatites at volatile-phase saturation (Lentz 1991). The absence of large cogenetic granite plutons in areas populated by uraniferous pegmatites indicates that parent magmas were small masses (Lentz 1991). Parent granite melts were derived by undersaturated partial melting of midcrustal rocks, perhaps the underlying Central Gneiss Belt (Lentz and Kretz 1989; Lentz 1991) due to crustal uplift following the Ottawan Orogeny (Lentz 1991). Consistent with this model is the generation of melts generated by frictional heating and fluid movement into crustal-scale faults (Lentz 1991). Faults and shear zones appear to have provided favourable structural sites for pegmatite emplacement; subsequent brecciation and shearing provided channelways for additional uranium deposition from late-stage hydrothermal fluids (Masson and Gordon 1981).

The general setting and timing of the late tectonic uraniferous and rare element-bearing pegmatites, skarns and fluorite-apatite-calcite veins is similar to that of the rare element and uraniferous pegmatites present in the Superior Province (*see* Fyon, Breaks et al., this volume) and a similar tectonic setting is implied. The Grenville Province deposits occur in a dominantly metasedimentary terrane, and are localized along regional structures. The Grenville Province deposits developed during crustal uplift, perhaps in response to tectonic thickening as a result of the integrated collision history of the Grenville Province (*see* Easton, this volume). This tectonic setting favoured the generation of fertile parental magmas from which the magmatic-hydrothermal fluids evolved, leading to the formation of uraniferous and rare element pegmatites. Skarns developed where the magmatic-hydrothermal fluids interacted with chemically reactive host rocks (Lentz 1991; Lumbers et al. 1990).

## METAL CONCENTRATION RELATED TO PRE–PALEOZOIC WEATHERING

Mineralization related to pre–Paleozoic weathering of Grenville basement includes: 1) supergene hematitization of Grenvillian magnetite skarn deposits; 2) concentration of gold and other metals in the regolith developed along the unconformity; 3) weathering of phlogopite-rich units to form vermiculite; 4) concentration of uranium and manganese along the unconformity; and 5) development of karst systems, with associated chalcopyrite and gold mineralization (DiPrisco and Springer 1991). Most of these mineralized areas represent concentrations of pre-existing Grenvillian mineralization (*see* Easton, this volume).

Two habits of hematite mineralization (type 9 deposits) occur: 1) disseminated to massive hematite within the Potsdam Sandstone at the base of the Paleozoic succession; and 2) hematite confined to fractures in Grenville marbles and Paleozoic limestone. Only minor production (less than 15 000 tonnes) of hematite has been achieved from these deposits (Carter 1984).

## POST–ORDOVICIAN VEINS

### Calcite-Barite-Fluorite-Strontium-Lead Veins

Calcite-barite-fluorite-strontium-lead veins (type 10 deposits) occupy late, vertical to nearly vertical faults and fractures which parallel northwest-striking post–Ordovician faults of the Ottawa–Bonnechere graben system (*see* Figure 24.9f; Carter 1984). Although these veins did not form in response to Grenville orogenic events, they cut Grenville and Ordovician rocks, and consequently are included in this section. Easton and Ford (1990) also reported lead anomalies along faults of the Ottawa–Bonnechere graben system where no veins are present. Similar anomalies are also visible along these fault traces on regional geochemical surveys (*see* Easton, this volume).

Coarse-grained, white to pink calcite is the principal vein mineral in most deposits, accompanied by varied amounts of barite, fluorite, galena, with locally abundant sphalerite and celestite (Carter 1984). These veins have been mined for fluorite, zinc and lead (e.g., Kingdon Mine; Carter 1984). The 3 most common mineral associations are: 1) calcite-fluorite-barite; 2) calcite-barite-galena; and 3) calcite-celestite (Sangster 1970; Sangster and Bourne 1982). Unlike the ca. 1.0 billion-year-old calcite-fluorite-barite veins of the Thunder Bay area (*see* Fyon, Bennett and Jackson, this volume), there is no associated mafic magmatism or precious metal mineralization.

### Calcite-Fluorite-Barite Veins

Fluorite-bearing veins almost all occur in the Madoc area in northwest-striking, steeply dipping fractures and fissures within, or associated with, the Deloro Granite (Easton 1989; Guillet 1964). The mineralized zones occur as connected lenses, disconnected *en échelon* lenses, or isolated occurrences within the fault zones. Individual lenses are up to 6 m thick, but average 3 m in thickness and 100 m in length (Guillet 1964). Host rocks include marble, granite, Ordovician limestone (Black River Group) and porcellanite. Veins can contain between 50 and 75% fluorite, 25% calcite, and 15% barite (Guillet 1964). Sphalerite is abundant at some deposits (e.g., Blakely deposit). The veins are vuggy and exhibit crustiform and comb structures. Brecciated wall rock fragments are common, as is banding (e.g., bands of pure fluorite up to 0.5 m thick and 26 m long occur at the Bailey deposit; Guillet 1964).

### Calcite-Barite-Galena Veins

Galena-bearing veins (e.g., Kingdon Mine and Crown King, Hollandia and Victoria deposits; Carter 1984) are the most common vein type and occur throughout southeastern Ontario, although they are most abundant along the Ottawa River. Calcite is the principal vein mineral, accompanied by either, or both, barite and galena, and minor sphalerite, marcasite, pyrite and chalcopyrite (Guillet 1964).

### Calcite-Celestite Veins

Calcite-celestite veins occur mainly in the southern Frontenac Terrane, and are mineralogically similar to the calcite-barite-galena veins, except for the increased abundance of celestite (Carter 1984).

### Discussion

The veins formed in an intracratonic tectonic setting in response to faulting along the post–Ordovician, Ottawa graben system (Carter 1984). The general lack of wall rock alteration associated with the veins and the fluid inclusion geothermometry data (McCartney 1964) indicate low temperature vein formation (Carter 1984). The close spatial relationship of the deposits to the Precambrian–Paleozoic unconformity, the simple mineralogy, open space filling textures, low temperature of deposition, and their geological setting and location near an intracratonic basin are consistent with the involvement of basinal or conate hydrothermal fluids (Carter 1984). Alternatively, they may have precipitated from meteoric hydrothermal fluids (Wilson 1929; Sangster 1970; Sangster and Bourne 1982).

## DISCUSSION

Few distinct periods of mineralization can be identified within the Grenville Province. Four periods of mineralization appear to have occurred in the Central Metasedimentary Belt, all associated with regional, or local, extensional regimes. 1) Between 1270 and 1240 Ma, most mineralization in the Central Metasedimentary Belt formed in association with volcanism, sedimentation and early plutonism (Killer Creek Gabbro, Lavant and Methuen suites), most likely in an extensional (marginal basin or back-arc) regime. 2) Between 1230 and 1180 Ma, epigenetic gold-quartz vein systems developed, perhaps coincident with the Elzevirian Orogeny, and minor sulphide mineralization associated with deposition of the Flinton Group. These latter 2 events may be related to the Frontenac Terrane docking with an Elzevirian Superterrane at ca. 1180 to 1160 Ma. Although this event was one of regional compression, local extension occurred along terrane and domain boundaries and in the Mazinaw Terrane, and that is where mineralization was localized. 3) Between 1050 and 1020 Ma, extension occurred throughout the Bancroft Terrane, leading to emplacement of granite pegmatites, molybdenum and uranium skarn deposits, and calcite-apatite veins. Some gold mineralization may have also occurred at this time along late extensional shear zones (e.g., Robertson Lake mylonite zone?). Finally, 4) between ca. 580 and 450 Ma, minor mineralization occurred in conjunction with Paleozoic sedimentation and the development of the Ottawa–Bonnechere graben system.

Although a variety of pluton types are present within the Central Metasedimentary Belt, showing a secular evolution from mafic to intermediate (gabbro-tonalite) to alkalic (syenite) compositions between 1350 and 1090 Ma (*see* Table 24.3), significant mineralization (copper-nickel, iron-titanium, industrial minerals) has only been identified in relation to the 1290 and 1245 million-year-old Killer Creek and Lavant suites, and the 1240 million-year-old Methuen Suite. Much of this mineralization is contact metamorphic, and the role of the specific magma is difficult to ascertain. Again, all 3 magma types appear to have been emplaced in extensional regimes; the gabbros are related to back-arc basin volcanism, whereas the granites show anorogenic affinities (Lumbers et al. 1990), possibly due to emplacement in a back-arc regime.

Even less is known about the history of the Central Gneiss Belt, in large part because of the paucity of known mineralization. Some events may be roughly coincident to events in the Central Metasedimentary Belt. Graphite deposits in the Algonquin Terrane are initially related back to sedimentation, although subsequent deformation and metamorphism is responsible for their quality and grade. Sulphide and magnetite deposits in the Parry Sound Domain may be associated with volcanism and sedimentation, and gold deposits in the same area might be coincident with ca. 1160 Ma metamorphism and deformation of the domain. A secular evolution in pluton types in the Central Gneiss Belt is less obvious than within the Central Metasedimentary Belt. Perhaps the lack of mafic to ultramafic and chemically evolved felsic plutonic suites within the Central Gneiss Belt is largely responsible for the limited plutonism associated mineralization in the Central Gneiss Belt.

## SUMMARY AND SUGGESTIONS FOR EXPLORATION

### Gold Deposits

At present, not enough is known about the timing, mineralogy, or emplacement history of most Grenville gold deposits to make meaningful comparisons with other gold depositional regimes (e.g., the Superior Province). Gold mineralization in the Central Metasedimentary Belt is related to localized extensional regimes that developed after consolidation of the metavolcanic-dominated terranes (analogous to greenstone belt amalgamation). No other direct analogies to Archean gold mineralization can presently be drawn. As noted above, factors controlling the siting of gold mineralization in the Grenville Province include: 1) occurrence near the greenschist-amphibolite facies transition hosted in greenschist facies rocks; 2) common host rocks are mafic metaigneous rocks or dolomitic marbles; 3) most occur in regional deformation zones, both within domains and at domain and terrane boundaries; 4) late trondhjemite dikes may occur proximal to the gold deposits; and 5) in the Mazinaw Terrane, there is a spatial association with the Mazinaw–Flinton group unconformity.

### Sulphide Deposits

Volcanic rocks within the Central Metasedimentary Belt, and perhaps some of the mafic gneisses within the Parry Sound Terrane, contain low to moderate potential for VMS mineralization. Most sulphide deposits exploited to date have been pyrite-pyrrhotite deposits. VMS potential appears to be highest in the Elzevir Terrane. Within the Grimsthorpe Domain, Cyprus-type deposits might be expected in the ca. 1300 to 1270 million-year-old tholeiitic sequences of the Canniff Complex and the Grimsthorpe Group. Within the Belmont Domain, the ca. 1250 million-year-old bimodal basalt-dacite-rhyolite suites, particularly those containing associated tholeiitic andesites (e.g., Belmont Lake volcanics and Turriff metavolcanics), have the highest VMS potential.

### Zinc Deposits

There is a spatial relationship between dolomitic marbles, the occurrence of stromatolites, and zinc deposits within the Central Metasedimentary Belt. Inferred depositional environments are always shallow water. Regional arsenic enrichments have been noted in conjunction with some deposits (e.g., Salerno Lake deposit, Easton 1987). Regional soil and till geochemical surveys have proven useful in deposit exploration (e.g., DiLabio et al. 1982). The largest deposits are localized in the Bancroft, Frontenac and Sharbot Lake terranes, all of which are dominantly carbonate terrains which may represent the partly preserved remnants of platformal and ramp-slope carbonate deposition. Evidence of evaporitic conditions (anhydrite, tourmalinites) is so far known only from the Balmat–Edwards deposit in New York State.

Potentially, 3 zinc deposit types may be present within the Central Metasedimentary Belt: sabkha, MVT and SEDEX types. The Balmat–Edwards deposit is the best example of a sabkha-type; the Flinton Group occurrences in the Mazinaw Terrane are the best example of the SEDEX type. Other deposits are not well enough known to classify according to processes. Exploration models for all 3 types are likely to be most effective when used in the Bancroft, Frontenac and Sharbot Lake terranes, in conjunction with regional geochemical surveys and the identification of shallow water, dolomitic and stromatolitic marble belts. Potential for finding additional zinc deposits is moderate. SEDEX type zinc mineralization potential is low to moderate in metasedimentary domains of the Algonquin Terrane.

### Platinum Group Element Deposits

Little systematic exploration for PGE deposits has occurred in the Grenville Province, hence the potential for such deposits is difficult to establish. PGE deposits in the Superior Province hosted in mafic intrusions are either associated directly with fractionation processes or assimilation of crustal material, or they are associated with late hydrothermal processes in the mafic intrusions (*see* Fyon, Breaks et al., this volume). Mafic intrusions of the Killer Creek Suite and Lavant Suite both show evidence of magmatic fractionation, primarily in the form of igneous layering, and minor PGE enrichments have been noted from Killer Creek Suite plutons (Wilson *in* Macdonald and Cherry 1988; Easton and Ford 1991). Lavant Suite plutons (e.g., Lavant and Mountain Grove gabbros) not only show evidence of igneous layering, but partly assimilated metaigneous inclusions suggest a complex fractionation process. In addition, many Lavant Suite plutons show some evidence of wall rock assimilation, particularly carbonate host rocks, as evidenced by intrusion breccias containing partly assimilated blocks, volatile-streaming in marginal rocks of the intrusions, and grain size and compositional heterogeneity in the margins of the intrusions. As a consequence, there may be a moderate potential for PGE deposits within the Lavant Suite and Killer Creek Suite in the Central Metasedimentary Belt. Within the Central Gneiss Belt, many of the ca. 1170 million-year-old metabasite bodies have been examined for their PGE potential, particularly in the Huntsville region (Garland and Villard 1990). No significant deposits have yet been reported.

## Industrial Minerals

Although a wide variety of industrial mineral commodities are present within the Grenville Province occurring in many different geologic settings, a few general statements can be made that may assist exploration for industrial minerals. First, many deposits are associated with the contact aureoles of Lavant Suite and Methuen Suite plutons (*see* Figure 24.10; e.g., talc, wollastonite, dolomite and tremolite). Second, productive graphite deposits occur dominantly within clastic or carbonate metasedimentary rocks within granulite facies terranes. An exception may be the Black Donald Mine in the Mazinaw Terrane, which is hosted in upper amphibolite facies rocks, however, the Mazinaw Terrane was subjected to 2 metamorphic events, and the earlier event may have locally attained granulite facies conditions. Third, aluminous metasedimentary rocks are not common within the Central Metasedimentary Belt. Most occur within the Flinton Group, and exploration for kyanite, sillimanite and mica deposits should focus on this stratigraphic unit in the Mazinaw Terrane. Fourth, deformation zones associated with terrane and domain boundaries throughout the Grenville Province are the best exploration targets for flagstone deposits. Finally, it should be emphasized that knowledge of the final use or market of the particular commodity is a necessary prerequisite to any industrial mineral exploration program. For example, dolomite used in the production of magnesium metal using the Pidgeon process must have low zinc contents (less than 100 ppm, Easton et al. 1986), whereas zinc content would be less important in uses as a mineral filler.

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## Chapter 25

# Tectonic Evolution of Ontario: Summary and Synthesis

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## Abstract

**Tectono-stratigraphic assemblages, of varied age and facies, are viewed as fundamental "building blocks" in the progressive development of greenstone belts, subprovinces and the Superior Province as a whole. Such assemblages are bounded by faults, unconformities and intrusions. They formed in specific sedimentary or volcanic environments, were progressively accreted into elongate greenstone belts over a period of approximately 300 million years up to about 2.7 Ga, underwent regional folding and thrust faulting, and were then invaded by volumetrically dominant tonalite-granodiorite masses. In late Neoarchean time (2.70 to 2.69 Ga), they acted as major source regions for huge intervening sedimentary basins. By 2.69 Ga, these had undergone tectonic contraction to form accretionary prisms, culminating in their coalescence with adjacent granite-greenstone subprovinces. At this time, Superior Province-wide shortening and crustal and magmatic thickening manifested as pervasive planar deformation fabrics and metamorphism, known as the Kenoran Orogeny.**

**Evolution of the Superior Province is considered directly comparable to that of younger orogens, in terms of rock type, structure and scale. Episodic production of supracrustal and plutonic rocks, and their aggregation as linear greenstone belts in elongate zones of intrusive and/or metamorphic activity, have parallels in younger arc-accretionary orogens. The geological development of the Superior Province is regarded as consistent with an accretionary plate tectonic model.**

**Postaccretionary crustal response included a rapid decrease in felsic magmatism in the cooling crustal mass, and extensive strike-slip faulting, fracturing and dike emplacement that continued spasmodically into Mesoproterozoic time.**

**The Archean cratons of North America are separated by linear belts of Proterozoic age. In Ontario, these belts are:**

1. **The Trans–Hudson Orogen (1.9 to 1.8 Ga) separates the Superior and Hearne provinces and underlies a small part of northwestern Ontario, largely beneath the Paleozoic and Mesozoic Hudson Bay Basin. This orogen includes volcanic rocks of the Fox River Belt and the passive margin to foredeep sequences of the Belcher fold belt.**
2. **The Penokean Orogen (2.4 to 2.2 Ga) separates the Superior Province, Yavapai Province and Mazatzal Belt and occurs along the shores of lakes Superior and Huron. The orogen consists of a passive margin sequence, the Huronian Supergroup, which is composed of 4 unconformity bounded siliciclastic units in the Lake Huron area. In the Lake Superior area, a continental margin to foredeep sequence is represented in Ontario by the Animikie Group.**
3. **The Mesoproterozoic Midcontinent Rift (1.1 Ga) consists of up to a 30 km thickness of plateau basalts and subordinate clastic sedimentary units in the Lake Superior area.**

**The Lake Superior region is underlain by minor anorogenic plutons, with associated volcanic rocks and the Mesoproterozoic epicontinental red beds of the Sibley Group.**

**The Grenville Province is a complex orogenic belt which truncates rocks of the Superior and Southern provinces, the Killarney Magmatic Belt, the Eastern Granite-Rhyolite Province and the Midcontinent Rift. The Grenville Province is divided into a northern part, the Central Gneiss Belt, and a southern part, the Central Metasedimentary Belt.**

**The Central Gneiss Belt is composed of highly metamorphosed parautochthonous gneissic and igneous rocks, which have been subdivided into distinct lithotectonic terranes and domains. The Central Gneiss Belt is made up of largely reworked Archean and Paleoproterozoic gneisses cut by Mesoproterozoic intrusions. Overlying and southeast of these parautochthonous terranes are several allochthonous terranes.**

**The Central Metasedimentary Belt is a major Mesoproterozoic accumulation of supracrustal units cut by plutons of several ages. The belt is subdivided from north to south into the Central Metasedimentary Belt Boundary Zone, and the allochthonous Bancroft, Elzevir, Mazinaw and Sharbot Lake terranes. The Central Metasedimentary Belt Boundary Zone is a major shear zone involving supracrustal and metamorphic rocks which may be largely allochthonous. The Bancroft, Elzevir, Mazinaw and Sharbot Lake terranes are the result of volcanism and sedimentation between 1300 and 1250 Ma, followed by plutonism and metamorphism at 1240 to 1250 Ma and 1140 to 1070 Ma. These terranes are characterized by low to medium metamorphic grade. The Frontenac Terrane to the south is a region of high metamorphic grade lacking volcanic rocks, but containing marbles, quartzites and quartzofeldspathic gneisses intruded mainly by granitoids and metamorphosed at 1170 to 1160 Ma.**

**The Grenville Orogen is a complex thrust stack assembled over the interval pre 1780 to 1000 Ma. The major orogenic events were the Elzevirian Orogeny (1240 to 1150 Ma) and the Grenville Orogeny (ca. 1150 Ma). These orogenic events completed the assembly of Laurentia, the North American craton (Hoffman 1989).**

**The stabilized Laurentian craton was subsequently overlain by a succession of Paleozoic and Mesozoic sedimentary rocks comprising 10 largely unconformity bounded depositional sequences. These sequences were largely deposited within the epicratonic Hudson Bay, Moose River and Michigan basins, and in the Appalachian Basin, a foreland basin. The intracratonic sedimentary basins of Ontario developed in regions of the craton well-removed from the influences of continental margin tectonism. However, the influence of distant tectonic events can be noted in the pulses of orogen-derived clastic sediments shed during the Taconic, Acadian and Alleghanian orogenies into the intracratonic basins. Cratonic tectonism, such as intracratonic basin subsidence and arch-centred uplift is noted in the shape, distribution and internal makeup of the depositional sequences within the intracratonic basins, and may have been driven by continental margin tectonism. In Ontario, intracratonic basins probably originated through the transmission of horizontal stress from the plate margin, causing subsidence, with location influenced by heterogeneities in the crystalline substrate of the basins. In Ontario, minor alkalic and kimberlitic igneous activity was associated with Paleozoic and Mesozoic tectonism. The rocks of intracratonic basins in southern Ontario have been displaced along minor faults, some of which involve Precambrian basement rocks. The presence of neotectonic structures and low-level seismic activity suggests that some of these faults may still be active.**

# PART 1: REVOLUTION IN THE SUPERIOR PROVINCE

**H.R. Williams, G.M. Stott and P.C. Thurston**

## INTRODUCTION

Part 1 of this chapter synthesizes earlier chapters of this volume and summarizes our current understanding of the development of the Superior Province during the Archean. It reviews imperfectly understood patterns of tectonic activity that are here interpreted to result in the progressive aggregation of greenstone assemblages into belts, and thence into subprovinces. Constraints on assemblage formation processes, timing and scale are discussed, using sketches of tectonic processes and environments to illustrate the temporal and spatial development of the province. Subsequent parts summarize the tectonic development of Proterozoic, Paleozoic and Mesozoic components of the Superior Province.

### Terminology

The first part of the chapter documents a recent structural, geochemical and geochronological revolution, in which the understanding of several purely descriptive, rigorously defined terms becomes critical to a new interpretation of the tectonics of the Superior Province. These terms are fully defined in the introduction to the Superior Province (*see* Thurston, chapter 4, this volume), but are here briefly reiterated.

**Terrane:** a fault-bounded package of strata that is allochthonous to, and has a geological history distinct from, the adjacent geologic units (e.g., Howell 1989, p.214).

**(Fabric) Domain** : a volume of rock bounded by compositional or structural discontinuities, within which there is structural homogeneity (cf., Bates and Jackson 1987).

**Tectonic Assemblage:** a package of stratified volcanic and/or sedimentary rock units built during a discrete interval of time in a common depositional or volcanic setting. It may be bounded by structural discontinuities, unconformities or intrusions. A terrane may comprise 1 or more assemblages (cf., Tipper et al. 1981; Gabrielse and Yorath 1989).

In Part 1 of this volume, greenstone belts were described commonly in terms of supracrustal assemblages forming a collage that developed over a protracted period. These assemblages were at first sporadically affected by intrusive magmatism, localized deformation and metamorphism, but were then subjected to province-wide deformation, magmatism and metamorphism, an event termed the Kenoran Orogeny. In these descriptions we preferred to use descriptive rather than genetic terminology. Many of the chapters did conclude with a genetic, tectonic interpretation of the data on the scale of an individual subprovince. In this chapter, we attempt to develop a universal (across the Superior Province) tectonic hypothesis to explain the data from all subprovinces. This hypothesis is based on recent interpretations of Phanerozoic orogens because we find remarkable similarity between modern and Archean tectonic patterns and scales.

### Modern Accretionary Orogens

Large-scale, plate tectonic interpretations of Phanerozoic orogens as ancient analogues of modern island arcs and related sedimentary basins initially appeared simple. It was gradually appreciated that in detail both ancient and modern orogens were far more complex, including disparate blocks within which there was lithological, chronological and structural homogeneity. Such blocks were called terranes by Coney et al. (1980).

Increasing use of the new nongenetic term "terrane" in complex Mesozoic–Cenozoic orogens of the Western Cordillera of North America (Howell et al. 1985) reflected a need to describe the lensoid, shear zone-bounded nature, and disparate character, age and history of myriad blocks within these orogens. A terrane was defined as a fault-bounded block, of any scale, comprising an area of rocks displaying structural, lithological and historical homogeneity. Evidence from other Phanerozoic orogens showed that terranes were commonly allochthonous relative to each other and to the ancient continental margin at which they were accreted (Keppie 1981; Williams and Hatcher 1982). Some indicate a migration of blocks over thousands of kilometres (Gabrielse and Yorath 1989). Terranes were shown to accrete to orogenic margins and become incorporated into them like flotsam against an obstruction. Some "superterranes" are aggregations of smaller units assembled prior to orogenic incorporation. Thus, orogens were largely a collage of accreted blocks, some of which are themselves a collage of disparate assemblages, including pieces of older sialic nucleii (Howell et al. 1985). With adequate geological and geophysical data, the complexities of modern orogens are explicable in terms of relatively simple basic "templates": examples of terrane accretion involving crustal components of diverse origin. These were recognized in orogens worldwide within a decade (Howell 1989).

In this chapter, we propose an accretionary origin for Archean rocks of the Superior Province and discuss similarities of scale and process between modern and ancient tectonics. We also invite comparison between greenstone assemblages and modern analogues, and seek to emphasize both similarities and differences.

## An Accretionary Model for the Superior Province

In this chapter we propose an accretionary mechanism as a viable, evolving interpretation of geological data on all scales. We consider that accretionary, as well as rift or wrench fault-related processes, are needed to explain the development of greenstone belts. For example, accretion adequately accounts for the diverse age ranges and paleoenvironments displayed and inferred in greenstone belts. Large-scale accretion, rather than rift or wrench-fault tectonics, best explains the evolution of sedimentary subprovinces (cf., Percival and Williams 1989; Williams 1990). Table 25.1 summarizes events in the Superior Province, and presents an overview of the episodic and repetitive nature of rock formation and deformation.

We will also document the evolution of much of the Superior Province by discussing the origin, development and assembly of its constituent lithotectonic components. The smallest, fundamental lithotectonic units are the tectonic assemblages described in detail in earlier chapters. It will be shown that volcanic sequence stacks have been tectonically juxtaposed to form the complex aggregations now referred to as greenstone belts. We will also show how data collected via classical stratigraphic methods initially conflicted with structural data, and how more sophisticated stratigraphic-structural analyses, allied to precise geochronological analyses, have radically changed the way geologists interpret the structure and formation of greenstone belts. Many belts contain older sequences lying beside or on top of younger rocks in configurations that are most readily explained by a stage of "thin-" or "thick-skinned" thrust faulting. Such belts, in varied stages of erosion, are separated by and interleaved with equant to elongate granitoid

**Table 25.1.** A simplified table of major tectonic events in the Superior Province, their timing, and the criteria used to distinguish them (*derived from* Charts A and B, Maps 2579 and 2580, map case); use in conjunction with Figure 25.7.

| Event | Time (Ma) | Criteria/Comment |
|---|---|---|
| Cratonization, continued thermal relaxation, brittle shearing, mylonite formation, mafic diking | post 2630? | Argon-argon blocking ages for hornblende 2.6 Ga and biotite 2.5 Ga; Matachewan diabase dike swarm 2.454 Ga; ages of mylonite and pseudotachylite 2.3 Ga |
| Hydrothermal alteration, associated alteration, brittle-ductile shearing and horizontal extension, retrogressive metamorphism, lamprophyre intrusion | 2630?–2670 | Ages of hydrous alteration in Hemlo |
| Late unconformable basins and alkalic magmatism, strike- and dip-slip faulting | 2660–2670 | Ages of Timiskaming volcanic rocks, and of clasts within coarse-grained sedimentary rocks |
| High-grade metamorphism, migmatization, genesis of granitic rocks, Kenoran deformation | 2670–2690 | Ages of syn- to posttectonic granitoids in greenstone subprovinces, pegmatites, S-type granites, metamorphic zircon grains in Ryan Migmatite and Kapuskasing Structural Zone |
| Calc-alkalic tonalitic magmatism in prisms, structural thickening in prism and at subprovince margins | 2688–2698 | Ages of plutons in Quetico Subprovince, ages of youngest volcanic rocks involved in subprovince accretion |
| Deposition of sediment in accretionary prisms, arc-coalescence | 2690–2698 | Zircon ages from Quetico and Pontiac sediments, upper limit |
| Final stage of calc-alkalic subaerial volcanic centres | 2698–2710 | |
| Episodic accretion of komatiitic, tholeiitic and diverse volcanic sequences; pre–Kenoran plutonism, deformation and metamorphism | 2700–28?? | |
| Mafic plain komatiitic and minor felsic volcanism, calc-alkalic plutonism | 2880–2889 | Age of Hawk assemblage |
| Calc-alkalic plutonism preserved in plutonic subprovinces and at margins of central part of Wabigoon Subprovince | 2830–3170 | Cedar Lake pluton 3003 ± 5 Ma<br>Marmion Lake batholith 2688 ± 3 Ma |

masses. Together, they constitute granite-greenstone terranes, or subprovinces.

In the late stages of greenstone belt development, granite-greenstone subprovinces became major sources of detritus which was transported into intervening, elongate sedimentary basins hundreds of kilometres in length. The aggregation of subprovinces at about 2.7 Ga induced a progressive deformation and plutonism which is known as the Kenoran Orogeny. After subprovince aggregation, the subprovinces acted as a relatively stable unit that reacted to stress by motion along subprovince boundaries or along new trans-subprovince faults.

While arguing that the Superior Province can reasonably be considered the product of arc accretion, we think it premature to offer small-scale explanations for the probably complex assemblage interactions that led to greenstone belt and subprovince development. Local descriptions and tectonic interpretations of parts of the Superior Province are presented in Part 1 of this volume (e.g., western Abitibi Subprovince, *see* Jackson and Fyon, this volume), but we caution against applying a detailed tectonic model for a subprovince (e.g., Uchi Subprovince as an Andean margin, *see* Stott and Corfu, this volume), or for the province as a whole. We are well aware of many outstanding problems of correlation and interpretation, and the inherent difficulties in solving these using conventional mapping and analytical techniques. Some outstanding problems are: •

1. We have few constraints on the small-scale structural processes which led to assemblage juxtaposition.
2. Within a subprovince, how do separate greenstone belts of contrasting metamorphic grade relate to each other, if at all?
3. Is the presence of old rocks (greater than 3.0 Ga) within most greenstone subprovinces evidence of fragmentation of an ancient greenstone subprovince, or did these old rocks develop independently in each subprovince?

## HISTORICAL REVIEW

Recently, there has been a fundamental change in perspective of the tectonic development of orogens (Coney et al. 1980). Multidisciplinary studies show that Phanerozoic orogens in part include discrete lithotectonic domains, or terranes, which are commonly bounded by ductile shear zones or faults. These terranes may be internally complex but can be distinguished by mutually contrasting geological histories and/or geotectonic environments. Relative motions between adjacent terranes may have been complex. Paleomagnetic and paleontological evidence indicates that terranes travelled both towards and parallel to the accretionary orogen (cf., Irving and Wynne 1991) for distances of up to several thousand kilometres. Accretionary orogens not only contain rocks that developed *in situ* but also rocks that are exotic. This conceptual revolution in turn prompted change in the perspective used to study Archean rocks, including those of the Superior Province.

### The Fixist Interpretation of Greenstone Belts

The classical fixist models for greenstone belts did not involve large-scale horizontal transport of greenstone belt units in a fashion comparable to that required in plate tectonic hypotheses. Formerly, most volcanic sequences in greenstone belts were thought to have developed in synformal wedges between buoyantly uplifted external batholiths, or in rifts within thin sialic crust (Anhaeusser et al. 1969). For example, the concept of mantled gneiss domes introduced by Eskola (1948) was used by MacGregor (1951) as a model for the development of gregarious batholiths in the Archean Rhodesian (Zimbabwe) craton.

Volcanic rocks were thought to have erupted in curvilinear basins on a pre-existing tonalitic gneiss-migmatite substrate: such basins evolved via gravitational instability (cf., Ramberg 1981) into synformal sinks between buoyant gneiss domes whose uplift eventually promoted the shedding of detritus back into the basins. One version of such a model invoked rift-related magmatism (Anhaeusser et al. 1969) and a resultant localization of volcanic basins along linear weak zones in a thin sialic crust. Such models explained the production, and structural and temporal control of mafic magmatism as an aspect of mantle behaviour, involving plumes or other upwellings (Campbell and Hill 1988). Temporal and chemical constraints on the generation and addition of felsic magma to higher crustal levels were rather unsatisfactorily explained by conductive heating consequent on crustal sagging or mafic underplating.

Worldwide, the classical tenets of greenstone belt geology included a tectonically undisturbed, albeit folded, layer-cake stratigraphy and cyclic volcanism within linear basins developed on a granitic substrate. Later, arc and back-arc depositional settings were widely advocated (Tarney et al. 1976). Some of these are discussed below, with examples from the Superior Province.

Vertically and laterally discrete greenstone belt volcanic successions were considered by many workers to be bounded by unconformities. Unconformities between supracrustal assemblages, or between plutonic and supracrustal rocks in individual greenstone belts, implied potential stratigraphic and structural complexity. Time gaps indicated by unconformities range up to 300 million years. Despite this, structural continuity of stratigraphic succession was rarely questioned. Data from early mapping (cf., Goodwin 1962) and a few relatively imprecise age determinations permitted a model of cyclic volcanism progressively younging upward in a stratigraphic succession that appeared to have synclinal form, especially in larger greenstone belts (e.g., Abitibi greenstone belt, MERQ/OGS 1983). Some lithostratigraphic successions in smaller greenstone belts seemed to have a homoclinal structure and were not readily explained using the classical synformal model (e.g., Sandy Lake greenstone belt, Sachigo Subprovince, Osmani and Stott 1988). Boundaries between cycles were commonly thought to be unconformities; evidence of fault contact was not recognized, or ignored in the face of established models (Nunes and Thurston 1980). Two corollaries derived from this recognition of apparently continuous successions: that estimated thicknesses for individual greenstone belts ranged up to 25 km (Jensen and Langford 1985) and that both individual and repetitive successions were largely autochthonous (Ayres and Thurston 1985).

## A Mobilist View of Greenstone Belts

Following the Krogh and Davis (1971) outline of a mobilist interpretation, Talbot (1973) challenged traditional fixist models in a seminal paper which presented a speculative model for Archean crustal development based on the new plate tectonic hypothesis. In the same year, Anhaeusser (1973) modified an earlier fixist model (Anhaeusser et al. 1969) to accommodate plate tectonics. Later, Tarney et al. (1976) compared Archean greenstone belts with parts of modern orogens such as the Rocas Verdes Complex in Chile, and a developing worldwide consensus began to recognize gross similarities between greenstone belts and the products of both continental and oceanic island arc environments. In the Superior Province, Langford and Morin (1976), Blackburn (1980) and Dimroth et al. (1983) developed more comprehensively field-based mobilist analyses of Archean geology. These form the basis of our present review, which is one of few dealing with most of the Superior Province (*see also* Card 1990).

Starting in the 1980s, several new techniques promoted rapid changes in the type of field and laboratory observations available to the geologist, including routine and precise trace element geochemical analysis, high-precision age determinations, structural analyses, and new ways to determine metamorphic pressures and temperatures. In concert, these produced a worldwide breakdown of old fixist models.

## Geochemistry

In the 1970s, the development of routine major and trace element analysis allowed a subdivision of greenstone belts into komatiitic, tholeiitic and mafic to felsic (calc-alkalic) volcanic units that were still interpreted as autochthonous cycles. Multielement, normalized plots using more alteration-resistant elements, such as the rare earth elements, titanium, phosphorus, yttrium, niobium, tantalum, thorium and zirconium, showed geochemical similarities between modern and Archean volcanic and plutonic rocks (cf., Barrie and Shirey 1991). Notwithstanding a caution by Hamilton (1988, p.1519), the new data led to discrimination of the tectono-depositional environments of volcanic rocks. For example, it was recognized that many massive to pillowed mafic lavas were compositionally similar to mid-ocean ridge basalt (MORB) tholeiites and that parts of the diverse mafic to felsic volcanic assemblages were physically and chemically indistinguishable from those in modern calc-alkalic arcs.

During this revolution, volcanologists recognized clear physical similarities between modern and ancient mafic to felsic volcanic complexes, as well as similar patterns of scale and alteration (cf., Dimroth et al. 1985). Indeed, perspectives gained from complete sections of ancient volcanic sequences aided the elucidation of modern volcanic processes. The assignment of greenstone belt rocks to different tectonic environments, that is, arcs, back-arc basins and ocean basins, and identification of their geochemical affinities (cf., Thurston and Chivers 1990), though often problematic, indicates that greenstone belts are not the result of a single process but a range of varied geological environments and processes (cf., Barrie and Shirey 1991).

## Precise $^{39}$Ar-$^{40}$Ar and U-Pb Age Determinations

During the 1970s, technical developments at the Royal Ontario Museum led to high-precision (±2 million year range) U-Pb zircon, baddeleyite and sphene age determinations of volcanic and intrusive greenstone belt rocks (Krogh 1982). Similar technical advances in K-Ar age determinations resulted in the development of high-precision Ar-Ar methods (York et al. 1981). Both methods produced rock formation age data, rather than data on the ages of metamorphism and deformation, and thus began to create chronostratigraphic problems.

The concept of a simple cyclic greenstone belt stratigraphy collapsed when geochronological data revealed old rocks overlying younger ones (Corfu et al. 1989). Blocks of ancient sialic crust with appended epicontinental cover successions began to be recognized, though their relationship to younger greenstone belts was equivocal. It was gradually realized that some greenstone belt successions were tectonically juxtaposed and that boundaries between them were not simple unconformities (cf., de Wit et al. 1987; Martyn 1988).

## Structural Geology and Tectonics

The new and rapidly developing recognition of chronostratigraphic complexity, based on newly refined techniques of isotopic age determination, implied tectono-stratigraphic complexity. Worldwide, geologists became more aware of the probable existence of tectonic boundaries between adjacent volcanic and sedimentary units in Archean greenstone belts. These contacts were probably shear zones but were frustratingly cryptic.

The same revolution that led to the recognition of large-scale shear zones at subprovince boundaries (e.g., Wawa–Quetico boundary, Stott 1986; Wabigoon–Quetico boundary, Williams 1986) also led to an understanding of shear zones and their kinematics at a greenstone belt scale (e.g., Michipicoten greenstone belt, Wawa Subprovince, Arias and Helmstaedt 1990). Schistose zones, previously mapped as tuffaceous rocks or ignored (e.g., Barberton greenstone belt, de Wit et al. 1987), are now readily recognized as occurring both within and marginal to volcano-sedimentary sequences (cf., de Wit 1991; *see* Stott and Corfu, this volume). Long considered unimportant, these hitherto intractable, poorly exposed, often highly weathered zones leapt in stature to become surfaces along which allochthonous cycles were emplaced (e.g., Beardmore–Geraldton Belt, Wabigoon Subprovince, Williams 1986).

Greenstone belts commonly contain contiguous sequences with opposed, back-to-back or front-to-front younging directions (e.g., Sandy Lake greenstone belt, Sachigo Subprovince, Thurston et al. 1987); these had hitherto been considered to result from folding, and the common presence of contrasting rock types on either "limb" was explained by convenient facies variation. Such sequences, and also aberrant cycles with sections apparently out of time sequence, have now been reinterpreted as tectonic artifacts (e.g., western Wabigoon Subprovince, Davis et al. 1988). The lithostratigraphic units constituting these sequences are tens of metres to several kilometres thick and were formerly regarded as tilted but otherwise undisturbed "layer-cake" strata, but are now considered to be examples of fault-bounded panels or duplexes such as those described by Swager and Griffin (1990).

Thus, the interpretive revolution in the Superior Province follows that in the Canadian Cordillera, which, by 1980 was viewed as made up of allochthonous slices (Gabrielse and Yorath 1989). Regarding the Superior Province, there was a change in perspective from autochthonous cyclicity to allochthonous juxtaposition of greenstone units; this led to a revitalization of ideas on how greenstone belts formed. Recognition and use of planar and linear kinematic fabrics, shear zones and strain estimates, allied with geochronological data, allowed the establishment of a provisional structural chronology, albeit from incomplete, irregularly collected data.

In many belts, the interpretation of kinematic structures and fabrics allowed the discrimination of 2 major deformational stages (Williams et al. 1989):

1. early shearing on shallowly dipping surfaces (e.g., Zimbabwe: Coward et al. 1976, Stowe 1974; Australia: Martyn 1988, Swager and Griffin 1990) and coeval recumbent folding (e.g., Superior Province: Arias and Helmstaedt 1990, Chorlton 1991).
2. later upright refolding (e.g., Lake of the Woods greenstone belt, Sanborn-Barrie 1991b), which in part had a spatial and temporal relation to both regional shortening across entire subprovinces and local pluton emplacement.

Greenstone belts, formerly thought to be simple synforms, now yielded evidence of a protracted, diverse stratigraphic and structural history. Greenstone belts in separate subprovinces apparently formed and deformed at different times, albeit with broad similarities in structural style (e.g., Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume; Wabigoon Subprovince, *see* Blackburn et al., this volume).

In the Superior Province, preliminary shear zone kinematic studies allowed the determination of slip vectors on major subprovince-bounding faults (e.g., Schwerdtner et al. 1978; Osmani and Stott 1988). Although few such shear zones have been thus studied, data indicate that most show complex movement directions, commonly starting with dip-slip vectors and ending with strike-slip vectors (e.g., Williams 1986; Robert 1989).

The strain distribution recorded in granite-greenstone subprovinces is very inhomogeneous. Early fabrics related to recumbent fold-thrust fault formation commonly predate granite emplacement (e.g., Michipicoten greenstone belt, Wawa Subprovince, Arias and Helmstaedt 1990; Sioux Lookout greenstone belt, Wabigoon Subprovince, Chorlton 1990) and indicate heterogeneously strained rocks. Greenstone belts were further affected by local or regional granite emplacement which produced strain aureoles of variable extent (e.g., Uchi Subprovince, *see* Stott and Corfu, this volume). Granite-greenstone subprovinces are traversed by shear zones, commonly located within and parallel to greenstone belts: these form curvilinear "strands" of highly strained rocks separating relatively weakly deformed lozenges. Strain intensity in granite-greenstone subprovinces is also a function of proximity to subprovince boundaries (e.g., Wabigoon Subprovince, *see* Blackburn et al., this volume).

## Metamorphism

Before the availability of precise temperature and pressure estimates, varying metamorphic grades in greenstone belts and in the intervening higher-grade subprovinces were interpreted as a function of differing volumes of pluton/batholith emplacement relative to greenstone belt rocks (Ayres 1978). However, recent metamorphic pressure estimates indicate great regional contrasts in erosion levels. These were locally inferred to result from differential uplift on faults (Percival and Card 1983) or differential crustal thickening (H.R. Williams 1991). For example, metamorphic pressure variation from 1 GPa (Percival 1983), to 0.2 to 0.3 GPa (Studemeister 1983) in eastern Wawa Subprovince, was ascribed to regional tilting related to crustal uplift along faults (Percival and Card 1983, 1985).

Metamorphic pressure estimates elsewhere in Superior Province suggest that apparent tilting of metamorphic isotherms, and differential erosion, need not be due solely to faulting (e.g., Uchi–Berens River subprovince transition, *see* Thurston et al., this volume). We argue that much, if not most, of the metamorphic pressure variation shown by presently exposed rocks is a primary tectonic and erosional response to areal variation in structural and/or magmatic crustal thickening.

# THE TERRANE CONCEPT APPLIED TO THE ARCHEAN

Terrane analysis in the Archean began several decades ago (e.g., Gill 1949; Douglas 1973; Stockwell 1982), with subdivision of the Superior Province into subprovinces, culminating in a review by Card and Ciesielski (1986). Many subprovinces are fault or shear-zone bounded, form continuous to discontinuous elongate masses, and display regular alternation of greenstone-rich and greenstone-poor regions on a scale of tens to hundreds of kilometres. Archean cratons other than the Superior Province contain contrasting tectonic units that are not as well developed, or on as large a scale. For example, Zimbabwe craton–Limpopo belt–Kaapvaal craton relationships in southern Africa may represent a similar style of alternation of granite-greenstone terrane with an intervening sediment-dominated prism as that documented from the southern Superior Province (Williams 1990).

Application of the concept of terrane analysis was for some geologists the next logical analytical step. This began as a nongenetic means of dealing with structural complexity, with the recognition of small tectonic units called assemblages within granite-greenstone subprovinces. This Special Volume marks the formal introduction of the term "assemblage" for rock packages in the Superior Province. Such relatively homogeneous lithological-structural-geochronological assemblages were distinguished from their neighbours on the basis of contrasting rock types and ages. Assemblage boundaries consist of faults, unconformities, or are indeterminate; some have later been "stitched" (dated plutons which straddle assemblage boundaries) together by plutons, and/or overlapped by supracrustal rocks or thrust sheets. The success of the approach is indicated by the fact that the use of such criteria resulted in the recognition of a finite number of assemblage types, some of which display contrasting structural styles (e.g., Jackson and Fyon 1991, this volume).

It became clear that subdivision of greenstone sequences into assemblages was applicable not only on the scale of a single greenstone belt, but also on a subprovince scale throughout the Superior Province. At the latter scale, groupings of assemblages are comparable in content and dimensions to those in modern orogens (e.g., Confederation and St. Joseph assemblages, southern Uchi Subprovince, *see* Stott and Corfu, this volume).

We carefully restrict our use of the term terrane; it is as yet undetermined everywhere in the Superior Province if, how and when assemblages accreted to form terranes on a regional scale. It seems that the term terrane is in part synonymous with usage of "subprovince", but we accept that some subprovinces may constitute aggregates of terranes as suggested for the Uchi Subprovince (*see* Stott and Corfu, this volume).

## ARCHEAN TECTONIC ASSEMBLAGES

Assemblages in the Superior Province are recognized largely on the basis of distinctive rock types and associations, structure and age. Such data may permit characterization of a lithostratigraphic environment, but we have yet to establish adequate discriminatory criteria for most environments, or clearly define boundaries between, for example, tholeiitic mafic plains and the mafic base of diverse mafic to felsic volcanic environments. The first tectonic assemblage map of the Canadian Cordillera delineated rock assemblages, each of which reflected "a unique environment or related environment of deposition or tectonic setting" (Tipper et al. 1981), but did not explicitly label such environments. The same approach is taken in the Tectonic Assemblages of Ontario map (*see* Maps 2575 to 2583, map case). Nonetheless, in the Superior Province, we do try to delineate contrasting lithostratigraphic environments such as epicontinental passive margin, fore-arc accretionary prism, komatiitic-tholeiitic mafic plain, back-arc basin, mixed mafic to felsic volcanic edifices, and late-stage unconformable fault-bounded basins (cf., Thurston and Chivers 1990).

Assemblages comprise rocks which are tectonically related in terms of environment of deposition, way-up, structural state, metamorphism and age. Assemblage boundaries are commonly faulted but may be unconformable (e.g., Wawa and Catfish assemblages), conformable, intrusive, or "stitched" by later intrusions (e.g., Trout Lake pluton, Uchi Subprovince). At the present stage of analysis, some assemblages are very large, incorporating an entire subprovince (e.g., Quetico assemblage, *see* Williams, this volume); some have great strike continuity (e.g., Confederation assemblage, Uchi Subprovince, *see* Stott and Corfu, this volume), whereas others cover only a few tens of square kilometres (e.g., Hawk assemblage, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume).

## ASSEMBLAGE TYPES

These are described in order of likely temporal appearance, to promote an appreciation of evolutionary development of the Superior Province (Table 25.2).

### Continental Platforms

These assemblages consist of thin clastic, chemical and biochemical rock sequences (Figure 25.1) formed on shallow water platforms (e.g., Sachigo Subprovince: North Spirit Lake, Nemakwis assemblage, Wood et al 1986; Favourable Lake greenstone belt, Ayres 1988). Recognition of such sequences, deposited on an older substrate, is recent (Wood et al. 1986; Thurston and Chivers 1990). At Steep Rock Lake in the Wabigoon Subprovince, clastic and biochemical sediments were deposited unconformably on granitoid rocks (*see* Blackburn et al., this volume); at Eyapamikama Lake in the northwestern Superior Province, clastic rocks unconformably overlie older volcanic rocks (*see* Thurston et al., this volume). Quartz-rich arenaceous detritus, the lithogical association of the assemblages, and deposition on lithologically variable crust may indicate epicontinental, tectonically inactive conditions. These assemblages are sporadically distributed: none are known in eastern Superior Province in Ontario.

### Oceanic to Continental Margin Assemblages

Syn-greenstone assemblages are, like continental platformal assemblages, developed prior to a stage of assemblage aggregation into greenstone belts, a process discussed later in this chapter. Late unconformable assemblages, discussed in the next section, are developed after, and commonly deposited unconformably upon (*see* Blackburn et al., this volume), syn-greenstone assemblages.

**Table 25.2.** Assemblage types within the Superior Province; Archean examples and comparison with modern examples of specific tectonic environments and structural units.

| Assemblage Type | Superior Province Examples | Phanerozoic to Modern Examples |
|---|---|---|
| Platform sediments | Steep Rock Group and sequence at Lumby Lake, Wabigoon Subprovince (*see* Blackburn et al., this volume); Keeyask metasediments, Sachigo Subprovince (Thurston and Chivers 1990; *see* Thurston et al., this volume) | Early part of platform cover on Yukon–Tanana terrane (Price and Hatcher 1983); Unazuki schist lying on pre-Permian Hida basement, Japan (Taira and Tashiro 1987) |
| Mafic plain | Lower volcanic sequence, Wabigoon Subprovince (*see* Blackburn et al., this volume); Hawk assemblage, Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume) | Mafic substrates to accreted sequences within Akiyoshi and Shimanto belts of Japan (Isozaki et al. 1990) |
| Volcanic arc | **Arc type:** Confederation assemblage, Red Lake, Uchi Subprovince (*see* Stott and Corfu, this volume); Catfish assemblage, Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume) | Wrangellia and Stikine terranes, North American Cordillera (Price and Hatcher 1983); southern Sierra Nevada (Saleeby 1990); Hidaka metamorphic belt, Japan (Taira 1985); Late Cretaceous volcanic rocks, Japan (Saito 1985; Koido 1991); Kohistan complex (Coward et al. 1987) |
|  | **Back-arc type:** Central assemblage, Sturgeon Lake, Wabigoon Subprovince (*see* Blackburn et al., this volume); Kidd–Munro assemblage, Abitibi Subprovince (*see* Jackson and Fyon, this volume) | Nikolai greenstone, Alaska (Beard and Barker 1988); Rocas Verdes Complex, Chile (Dalziel 1986); Miocene "green tuff" of Japan Sea (Zhang 1985) |
|  | **Fore-arc type:** Beardmore–Geraldton Belt, Wabigoon Subprovince (*see* Williams, this volume; *see also* Blackburn et al., this volume); Pickle Crow and Bruce Channel assemblages, Uchi Subprovince (*see* Stott and Corfu, this volume) | Shimanto Belt, Japan (Needham and Mackenzie 1988; Osozawa 1988) |
| Sediment prism | **Fore-arc type:** Quetico assemblage (*see* Williams, this volume); English River assemblage (*see* Breaks, this volume) | Kodiak Formation, Alaska (Sample and Moore 1987); Southern Uplands (Leggett et al. 1983); Dunnage Zone (van der Pluijm 1986); Shimanto Belt (Taira and Tashiro 1987) |
|  | **Intra-greenstone turbidite association:** Porcupine assemblage, Abitibi Subprovince (*see* Jackson and Fyon, this volume); Gamitagama assemblage, Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume); Billet assemblage, Uchi Subprovince (*see* Stott and Corfu, this volume) | Mino-Tamba belt, Japan (Isozaki et al. 1990) |
| Late, fault-bounded and unconformable basins | Shebandowan and Knife Lake assemblages, Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume); Timiskaming assemblage (*see* Jackson and Fyon, this volume); Bijou Point sedimentary rocks (*see* Thurston et al., this volume) | Fig Lake graben, North American Cordillera (Thorkelson 1989); others in British Columbia (Eisbacher 1985); Izumi basin, Japan (Taira and Tashiro 1987) |
| Greenstone belt | **Monocyclic:** Manitouwadge–Hornepayne assemblage, Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume) | Rocas Verdes Complex, Chile (Dalziel 1981) |
|  | **Polycyclic:** Confederation assemblage, Uchi Subprovince (*see* Stott and Corfu, this volume) | Southern Sierra Nevada (Saleeby 1990) |
| Subprovince | Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume) | Stikine terrane, North American Cordillera (Jones et al. 1986); Cordilleran and Appalalachian orogens (Williams and Hatcher 1982; Price and Hatcher 1983) |
| Province | Superior Province (this volume) | North American Cordillera (Coney et al. 1980; Price and Hatcher 1983) |

Greenstone belts include supracrustal sequences and/or assemblages developed in 2 contrasting volcanic regimes:

1. Mafic plain assemblages consist of komatiitic and tholeiitic basalts (e.g., Balmer assemblage, Uchi Subprovince, *see* Stott and Corfu, this volume) intercalated with deep-water sedimentary rocks such as sulphidic argillite. In the Abitibi Subprovince, Jackson and Fyon (this volume) have recognized distinct assemblages made up of komatiites, felsic volcanic rocks and iron formation. Volcanologic evidence suggests that these rocks are ancient analogues of modern oceanic, fore-arc or back-arc volcanism (*see* Jackson and Fyon, this volume; Figure 25.2); the limited geochemical data available support comparison with recent oceanic rocks (Thurston 1990a).
2. Island arc assemblages consist of calc-alkalic mafic to felsic volcanic rocks (e.g., Confederation assemblage, Uchi Subprovince, *see* Stott and Corfu, this volume). These probably represent ancient, primitive to evolved continental (Sylvester et al. 1987), island arc (Barley et al. 1989) and back-arc regimes (e.g., Kidd–Munro assemblage, Abitibi Subprovince, *see* Jackson and Fyon, this volume). Figure 25.3 shows an example of distribution and age ranges of major volcanic assemblages in part of the Uchi Subprovince.

## Syn- to Postaccretionary Sedimentary Assemblages

These assemblages are interpreted to have been deposited in varied inferred volcano-sedimentary apron environments, some of which are depicted in Figure 25.4. They include:

1. Proximal epiclastic sedimentary facies are represented by breccia and conglomerate (e.g., Doré conglomerate,

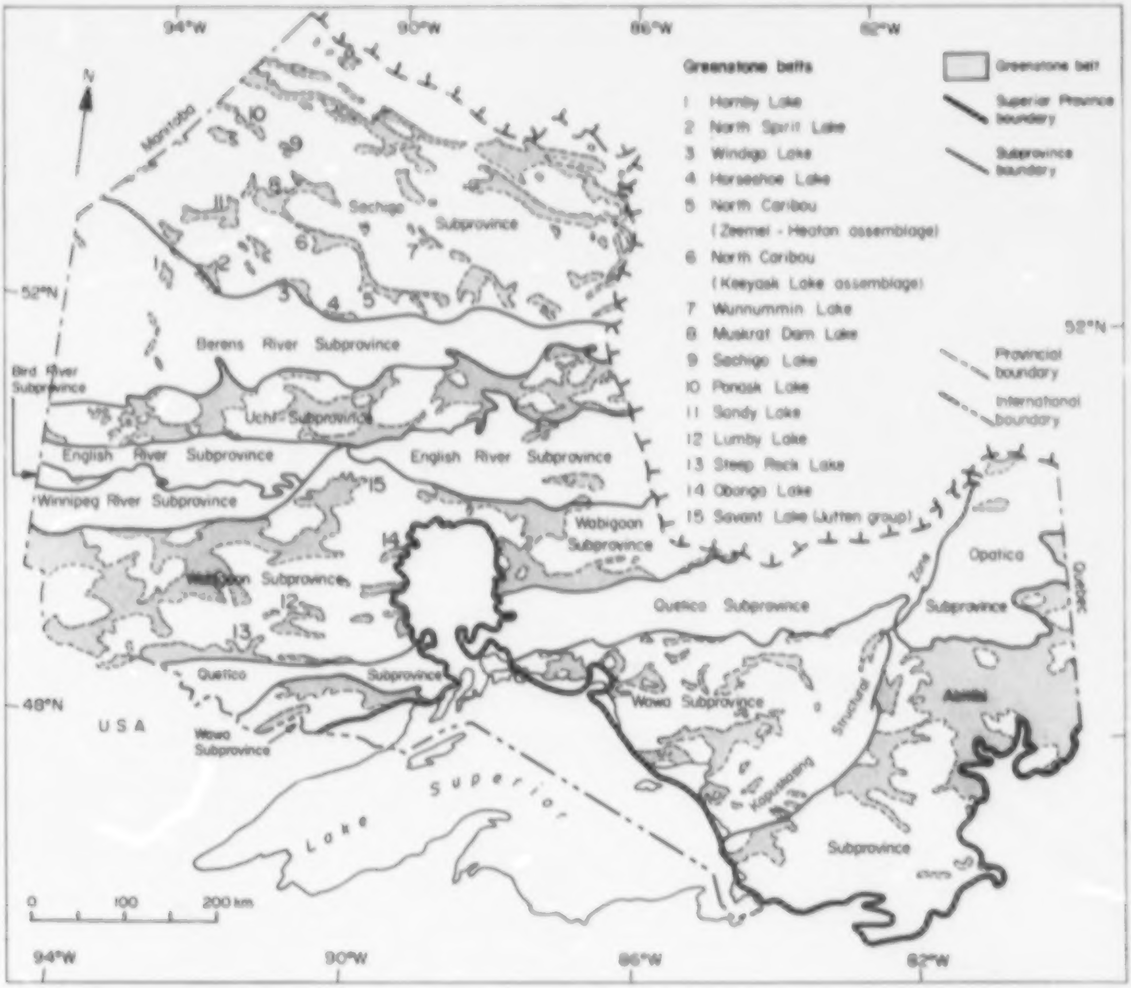

94°W
90°W
86°W
82°W
N
Greenstone belts
1 Hornby Lake
2 North Spirit Lake
3 Windigo Lake
4 Horseshoe Lake
5 North Caribou (Zeemel - Heaton assemblage)
6 North Caribou (Keeyask Lake assemblage)
7 Wunnummin Lake
8 Muskrat Dam Lake
9 Sachigo Lake
10 Ponask Lake
11 Sandy Lake
12 Lumby Lake
13 Steep Rock Lake
14 Obonga Lake
15 Savant Lake (Jutten group)
Greenstone belt
Superior Province boundary
Subprovince boundary
Provincial boundary
International boundary
52°N
48°N
Manitoba
Sachigo Subprovince
Berens River Subprovince
Bird River Subprovince
Uchi Subprovince
English River Subprovince
English River Subprovince
Winnipeg River Subprovince
Wabigoon Subprovince
Wabigoon Subprovince
Quetico Subprovince
Quetico Subprovince
Wawa Subprovince
Wawa Subprovince
Opatica Subprovince
Abitibi Subprovince
Quebec
Kapuskasing Structural Zone
USA
Lake Superior
0 100 200 km

**Figure 25.1.** Geological map of the western part of the Superior Province showing major tectonic subdivisions and greenstone belts, which include quartz-rich epicontinental sedimentary rocks developed upon a pre-existing sialic or mafic volcanic substrate. Details of assemblage descriptions are found in earlier chapters. Adapted from the Bedrock Geology of Ontario map (*see* Maps 2541 to 2545, map case).

Doré assemblage, Wawa Subprovince, Goodwin 1962; *see also* Williams, Stott, Heather, et al., this volume).

2. Wacke, siltstone and iron formation formed in basins on a scale of tens to hundreds of kilometres (e.g., Quetico assemblage, Quetico Subprovince, *see* Williams, this volume; Ryan Migmatite and Gargantua assemblage, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume).
3. Intra-greenstone belt successions of wacke, siltstone and minor iron formation (e.g., Whitney assemblage, Abitibi Subprovince, *see* Jackson and Fyon, this volume; Billett assemblage, Uchi Subprovince, *see* Stott and Corfu, this volume; Mishi assemblage, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume). Some may be equivalent to sedimentary wedge assemblages at faulted boundaries between volcanic assemblages, for example, shear-bounded lozenges in Quebec (Hubert 1990); others may be late examples of continental exhumation, unconformably straddling several volcanic assemblages (e.g., Billett assemblage, Uchi Subprovince, *see* Stott and Corfu, this volume). Intra-greenstone belt sedimentary assemblages are generally less than 1 km thick and separate volcanic assemblages of differing age and facies. The distribution shown in Figure 25.4 is typical of many greenstone belts: such sedimentary rocks are not always "axial" (as in classical models, e.g., Anhaeusser et al. 1969) but may form edges of belts and are commonly related to major shear zones.

## Late Unconformable Basins

Conformable to unconformable basins of mainly alkalic volcanic and fluviatile sedimentary rocks (Figure 25.5) formed late in greenstone belt history (Mueller et al., in press), and were historically termed "Timiskaming type". The best examples are the Timiskaming assemblage, Abitibi

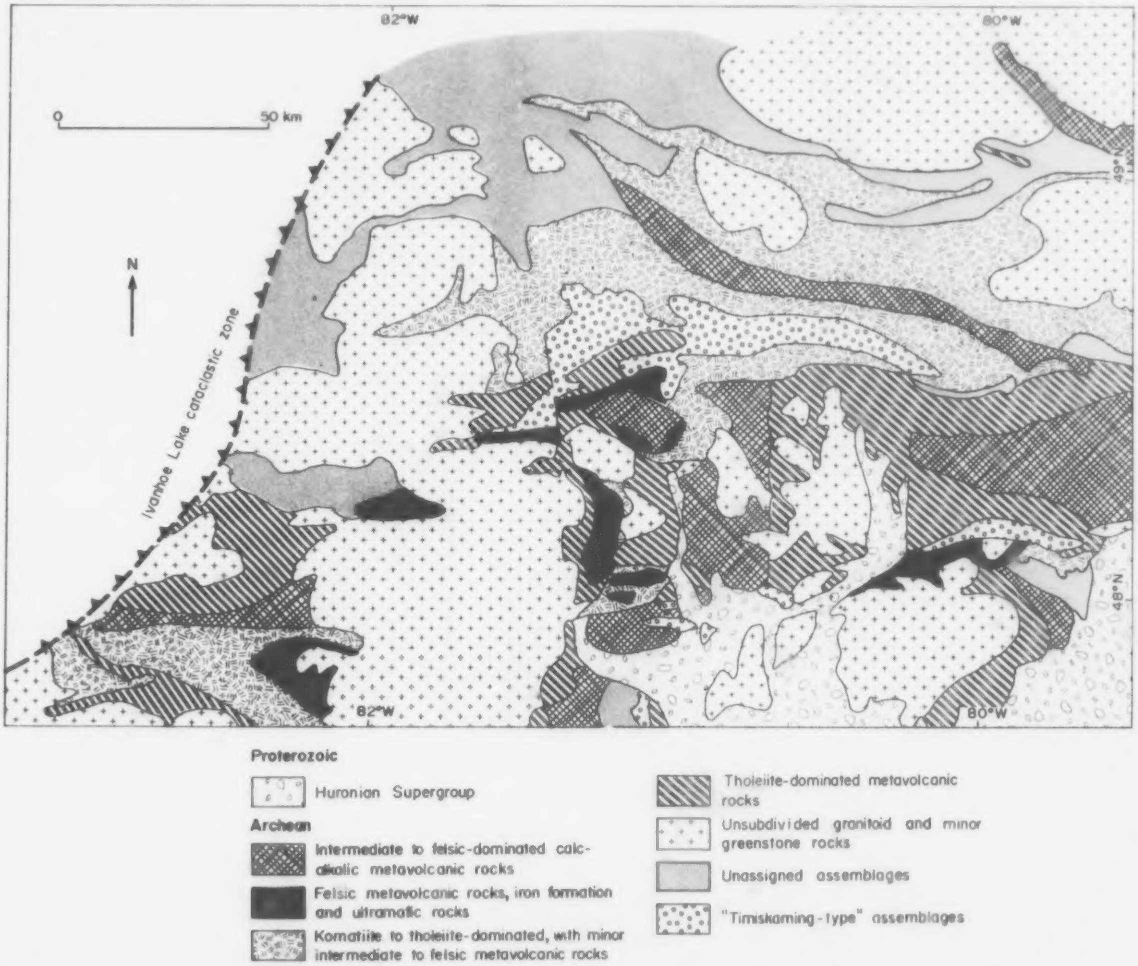

82°W
80°W
0
50 km
N
49°N
48°N
Ivanhoe Lake cataclastic zone
82°W
80°W
Proterozoic
Huronian Supergroup
Archean
Intermediate to felsic-dominated calc-alkalic metavolcanic rocks
Felsic metavolcanic rocks, iron formation and ultramafic rocks
Komatiite to tholeiite-dominated, with minor intermediate to felsic metavolcanic rocks
Tholeiite-dominated metavolcanic rocks
Unsubdivided granitoid and minor greenstone rocks
Unassigned assemblages
"Timiskaming-type" assemblages

**Figure 25.2.** Tectonic assemblage map of the western part of the Abitibi Subprovince depicting Neoarchean komatiitic, tholeiitic and calc-alkalic volcanic assemblages (*modified from* Figure 11.35, this volume).

Subprovince (Cooke and Moorhouse 1969; Hyde 1980; *see* Jackson and Fyon, this volume), and the Shebandowan assemblage, Wawa Subprovince (Shegelski 1980; *see* Williams, Stott, Heather et al., this volume).

## ASSEMBLAGE BOUNDARIES

Shear zones, unconformities and intrusive contacts can form assemblage boundaries; however, most boundaries are presently indeterminate and may be faults or unconformities. In some cases, a boundary in a volcanic sequence is inferred at the approximate line separating oppositely facing strata.

The history of movement along sheared assemblage boundaries is commonly complex. Such boundaries are recognized as high-strain zones on the basis of deformed markers and/or the development of a strong schistosity. The amount of offset between adjacent assemblages is usually undetermined, due to a lack of suitable markers. It can be argued that offset represents late, assemblage post-juxtaposition displacement and is unrelated to the pre-juxtaposition travel distance of an assemblage. Examples of shear-bounded assemblages occur in all granite-greenstone subprovinces (e.g., Boston, Hoyle and Kamiskotia assemblages, Abitibi Subprovince, *see* Jackson and Fyon, this volume; Balmer and Confederation assemblages, Uchi Subprovince, *see* Stott and Corfu, this volume).

Well-documented unconformities that are also assemblage boundaries are relatively rare. Some of the best-observed are those between volcanic assemblages and late unconformable greenstone assemblages (e.g., Timiskaming assemblage, *see* Jackson and Fyon, this volume; Shebandowan assemblage, *see* Williams, Stott, Heather et al., this volume). They also occur between volcanic units of similar ages (e.g., Catharine–Pacaud and Skead assemblages, Abitibi Subprovince, *see* Jackson and Fyon, this volume) and between sialic substrate and volcanic assemblages (e.g., Marmion Lake batholith and Lumby assemblage, Steep Rock Group, Wabigoon Subprovince, *see* Blackburn et al., this volume; Agutua Arm and Keeyask assemblages, North Caribou Lake, *see* Thurston et al., this volume).

Intrusive contacts are common assemblage boundaries. Most assemblages, especially in small greenstone belts, are partly bounded by an intrusive granitoid body, though contacts may also be sheared (e.g., Manitouwadge–Hornepayne greenstone belt, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume). Dated plutons which straddle assemblage boundaries ("stitching" plutons) constrain the age of assemblage juxtaposition (e.g., Burchell pluton, Wawa Subprovince, Corfu and Stott 1986; *see also* Williams, Stott, Heather et al., this volume). They may also provide a minimum age for an assemblage (e.g., Blalock pluton of 2688 Ma age intruding the Quetico assemblage, Davis et al. 1990) or a maximum age (e.g., Marmion Lake batholith of 3004 Ma age underlying assemblages in the Lumby Lake greenstone belt, Davis and Jackson 1988).

## AGE RANGES WITHIN AND BETWEEN ASSEMBLAGES

The use of precise U-Pb age determinations to establish the range of rock formation ages in an assemblage is limited by a lack of suitable material. Mafic rocks can rarely be dated precisely; felsic rocks, which can be dated, commonly occur at or near the stratigraphic top of an assemblage. Upper age limits of many assemblages are well constrained but the age of the oldest supracrustal material is usually poorly constrained. For example, the ~2.70 Ga upper limit of the Catfish assemblage (Wawa Subprovince) has been constrained by the age of felsic volcanic rocks near the top of the assemblage and by the felsic intrusions which cut it; the lower limit is constrained only by the ~2.75 Ga age of felsic rocks in the unconformably underlying Wawa assemblage. At Red Lake (Uchi Subprovince), the oldest basalts have been dated at 3070 Ga (Gulson et al. 1990) whereas the oldest felsic rocks have been dated at 2980 Ma (Corfu and Wallace 1986).

The age ranges used for the tectonic assemblage map were specifically chosen to minimize cases of an assemblage apparently straddling 2 time periods. It was hoped that careful choice of time intervals would allow the recognition of periods during which supracrustal development was at a minimum, thus indirectly recognizing the timing of orogenic motions during the aggregation of assemblages or groups of assemblages. Assemblages consisting of rocks which span a short age range (e.g., 2750 to 2730 Ma or 2710 to 2695 Ma) have been found in most subprovinces; some include well-constrained units with age ranges of only 10 to 20 million years . One result of our inability to determine if rocks in greenstone belts are allochthonous, or not, is that the reality of large-scale autochthonous volcanic episodicity in a greenstone belt is not well established (*see* Corfu and Davis, this volume).

Adjacent assemblages may show little difference in age range or relative age constraints (e.g., assemblages of 2740 to 2720 Ma age, Confederation and St. Joseph assemblages, Uchi Subprovince, Central Sturgeon Lake volcanics within the Wabigoon assemblage, Wabigoon Subprovince) or have markedly different ages (e.g., Wawa and Hawk assemblages of 2.75 and 2.89 Ga age, respectively, Wawa Subprovince; assemblages of 2.9 Ga, 2.8 Ga and 2.7 Ga age, Uchi Subprovince). Table 25.1 summarizes time-space relationships in the Superior Province and Figure 25.3 shows an example of assemblage distribution in central Uchi Subprovince.

## USE OF PRECISE AGES FOR COMPARISON OF ASSEMBLAGES

Greenstone belt assemblages of similar lithological and geochemical character may differ in age, either markedly (e.g., Hemlo and Heron Bay assemblages, Wawa Subprovince; Woman and Balmer assemblages, Uchi Subprovince), or slightly (e.g., Confederation and St. Joseph assemblages, Uchi Subprovince). Without age determinations, it is generally impossible to distinguish adjacent, lithologically similar assemblages. For example, the Hemlo

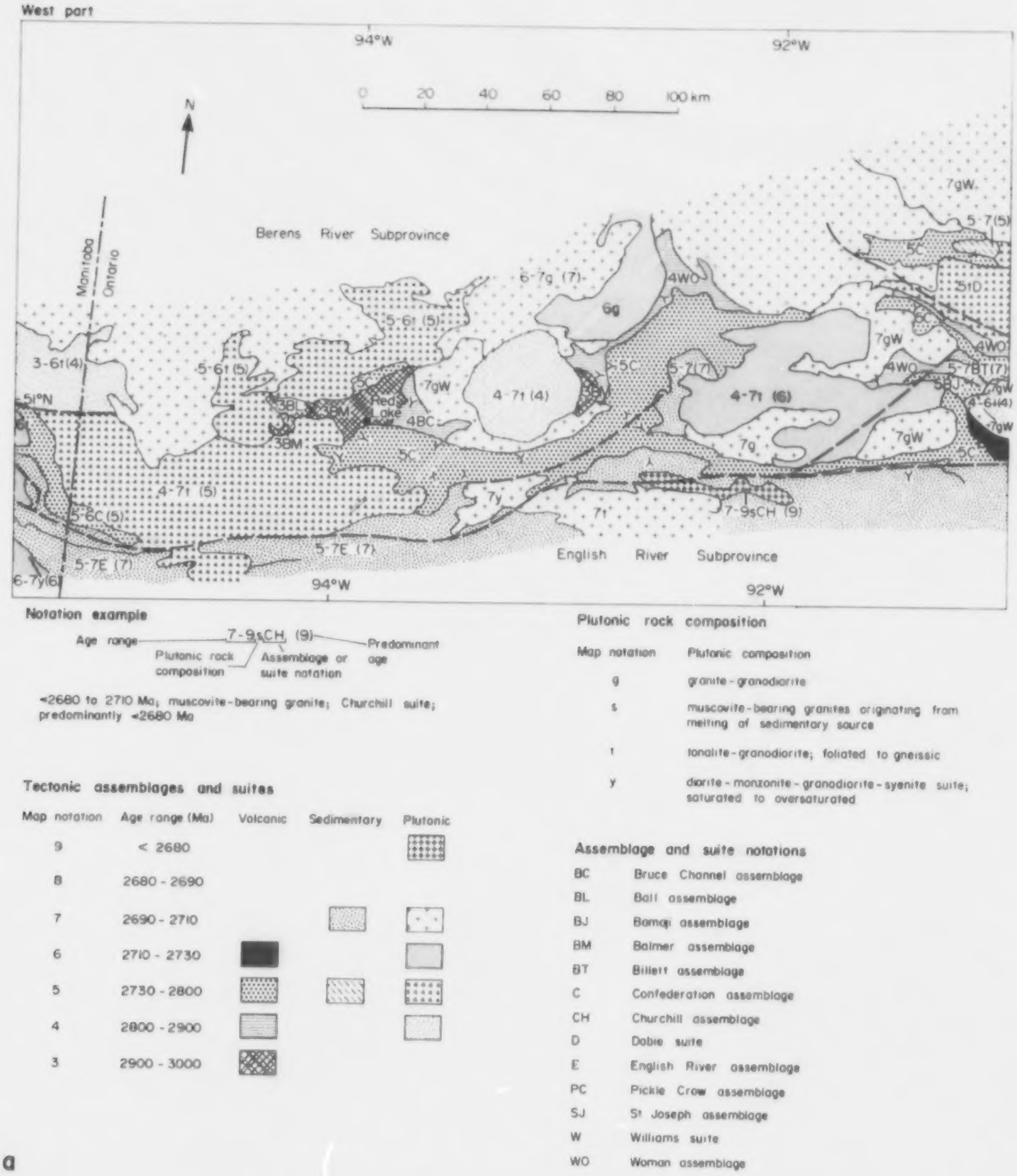

West part
94°W
92°W
0 20 40 60 80 100 km
N
Berens River Subprovince
Manitoba
Ontario
3-6t(4)
51°N
5-6t (5)
5-6t (5)
6-7g (7)
6g
4WO
3BL
3BM
3BM
Red Lake
4BC
7gW
4-7t (4)
5C
5-7(7)
4-7t (6)
7gW
7gW
4WO
5-7BT(7)
5BJ
4-6t(4)
7gW
5C
5D
5C
5-7(5)
7gW
5C
7g
4-7t (5)
5-6C(5)
5-7E (7)
5-7E (7)
6-7y(6)
5C
7y
7t
7-9sCH (9)
English River Subprovince
94°W
92°W
Notation example
Age range
7-9sCH (9)
Predominant age
Plutonic rock composition
Assemblage or suite notation
<2680 to 2710 Ma; muscovite-bearing granite; Churchill suite; predominantly <2680 Ma
Plutonic rock composition
Map notation
Plutonic composition
g
granite - granodiorite
s
muscovite-bearing granites originating from melting of sedimentary source
t
tonalite-granodiorite; foliated to gneissic
y
diorite - monzonite - granodiorite - syenite suite; saturated to oversaturated
Tectonic assemblages and suites
Map notation
Age range (Ma)
Volcanic
Sedimentary
Plutonic
9
< 2680
8
2680 - 2690
7
2690 - 2710
6
2710 - 2730
5
2730 - 2800
4
2800 - 2900
3
2900 - 3000
Assemblage and suite notations
BC
Bruce Channel assemblage
BL
Ball assemblage
BJ
Bamaji assemblage
BM
Balmer assemblage
BT
Billett assemblage
C
Confederation assemblage
CH
Churchill assemblage
D
Dobie suite
E
English River assemblage
PC
Pickle Crow assemblage
SJ
St Joseph assemblage
W
Williams suite
WO
Woman assemblage
a

**Figure 25.3.** Distribution and age of volcanic and sedimentary assemblages, and plutons, in the central part of the Uchi Subprovince. **a)** Map of tectonic assemblages, from Red Lake to Pickle Lake. **b)** North-trending cross section of the Uchi Subprovince near the Uchi–English River subprovince boundary (position of section line is shown on **a**; data have been projected along strike onto the section). Notations and age subdivisions used here correspond to those used on the Tectonic Assemblages of Ontario map (*see* Maps 2575 to 2578, 2583, map case).

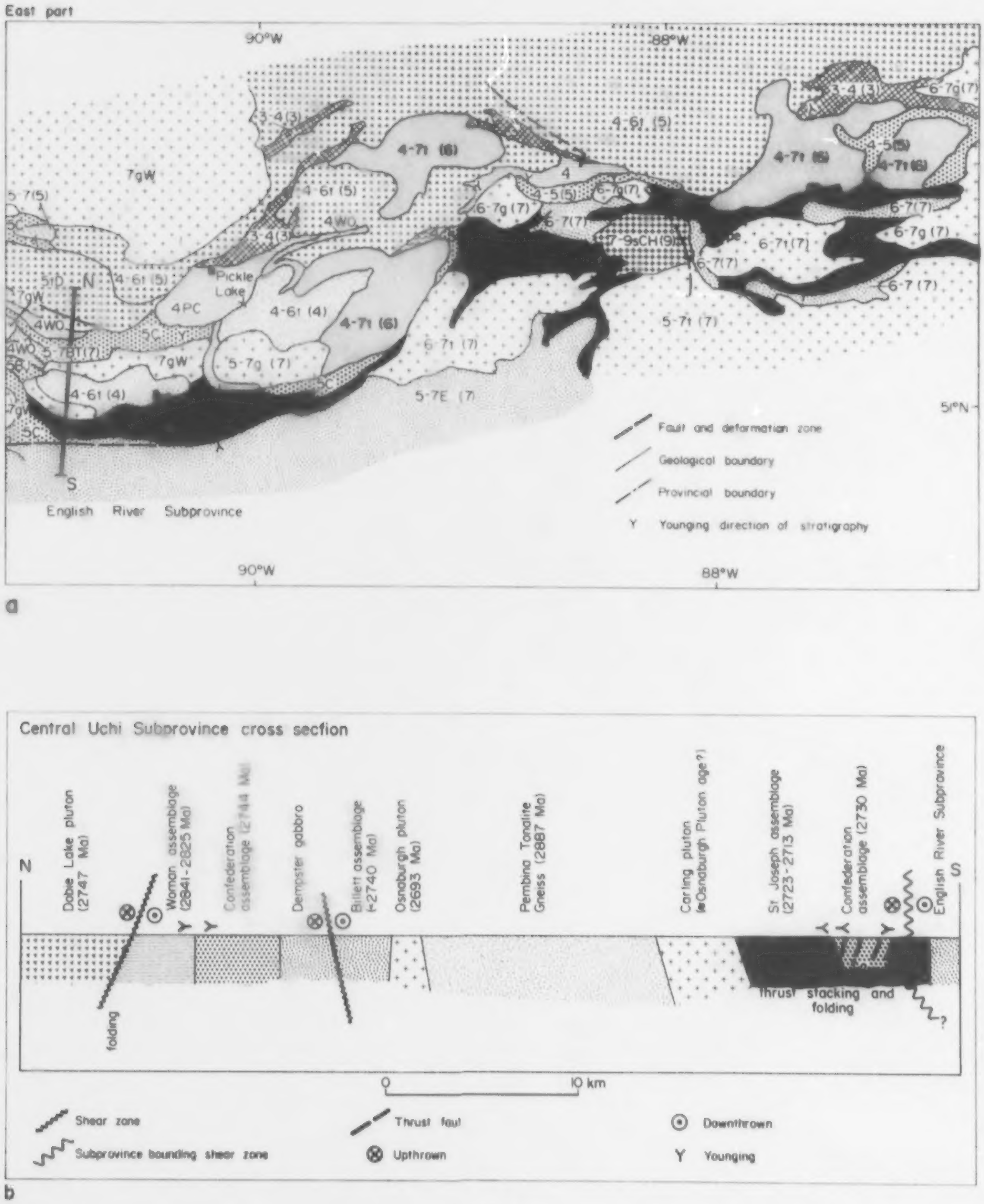

East part
90°W
88°W
3-4(3)
7gW
4-6t (5)
4-7t (6)
4
5(5)
6-7g(7)
6-7g (7)
6-7(7)
9sCH(9)
6-7t(7)
6-7 (7)
5-7(5)
5C
4WO
5ID
N
4-6t (5)
Pickle Lake
4PC
4-6t (4)
5-7BT(7)
7gW
5-7g (7)
6-7t (7)
5-7t (7)
5-7E (7)
5B
4-5(5)
6-7g(7)
51°N
S
English River Subprovince
Fault and deformation zone
Geological boundary
Provincial boundary
Y Younging direction of stratigraphy
a
Central Uchi Subprovince cross section
N
Dobie Lake pluton (2747 Ma)
Woman assemblage (2841-2825 Ma)
Confederation assemblage (2744 Ma)
Dempster gabbro
Billett assemblage (~2740 Ma)
Osnaburgh pluton (2693 Ma)
Pembina Tonalite Gneiss (2887 Ma)
Carling pluton (=Osnaburgh Pluton age?)
St. Joseph assemblage (2723-2713 Ma)
Confederation assemblage (2730 Ma)
English River Subprovince
S
folding
thrust stacking and folding
?
0
10 km
Shear zone
Subprovince bounding shear zone
Thrust fault
Upthrown
Downthrown
Younging
b

**Figure 25.3.** Distribution and age of volcanic and sedimentary assemblages, and plutons, in the central part of the Uchi Subprovince. **a)** Map of tectonic assemblages, from Red Lake to Pickle Lake. **b)** North-trending cross section of the Uchi Subprovince near the Uchi–English River subprovince boundary (position of section line is shown on **a**; data have been projected along strike onto the section). Notations and age subdivisions used here correspond to those used on the Tectonic Assemblages of Ontario map (*see* Maps 2575 to 2578, 2583, map case).

greenstone belt (Muir 1982) was considered a single lithotectonic unit until age data indicated an age range of 2772 to 2695 Ma within the belt. Elsewhere, experience suggests that such an age range generally spans more than 1 assemblage. No lithological basis for differentiating the northern and southern parts of the Hemlo greenstone belt was recognized during mapping, though there is a structural discordance along the axis of the belt. This feature is now used, in concert with geochronological data, to demarcate the boundary between 2 assemblages of contrasting age. In general, the volcanic activity represented by individual assemblages in greenstone belts rarely seems to span an age range of more than 20 million years; sequences with an age range greater than this commonly include rocks formed in differing depositional environments or contain a tectonic contact between parts of the sequence (e.g., Boyer and Wabigoon assemblages, Wabigoon Subprovince, *see* Blackburn et al., this volume; Krist, Boston and Catharine–Pacaud assemblages, Abitibi Subprovince, *see* Jackson and Fyon, this volume).

In some cases, there are similarities in assemblage ages in different greenstone belts in the same subprovince. In the Wawa Subprovince, for example, the Heron Bay and Catfish assemblages both yield ages of 2700±5 Ma. Slightly differing ages between noncontiguous assemblages may be significant: either the age difference is significant and the assemblages should not be correlated but will be distinguishable when additional data become available, or there are diachronous changes in age range of similar tectonostratigraphic units across or along subprovinces. We cannot yet choose between these alternatives.

Failing evidence to the contrary, we consider that individual granite-greenstone subprovinces evolved as separate entities, a thesis supported by contrasting initial $^{206}Pb/^{207}Pb$ and $^{143}Nd/^{144}Nd$ ratios from one subprovince to another (e.g., Shirey and Carlson 1988). Thus, we have not tried to correlate lithotectonic assemblages of similar age in different subprovinces, as for example volcanic rocks of 2730 Ma age in Uchi, Wabigoon and Wawa subprovinces.

## ASSEMBLAGE DEFORMATIONAL STYLES

Fold styles and strain intensity in greenstone belt supracrustal rocks are highly variable. There is great inhomogeneity of deformation within and between belts and from one subprovince to another. Strain variation is spatially related not only to variation in fold intensity, but also to pluton emplacement and expansion, and to shear zones, especially those at assemblage contacts and subprovince margins.

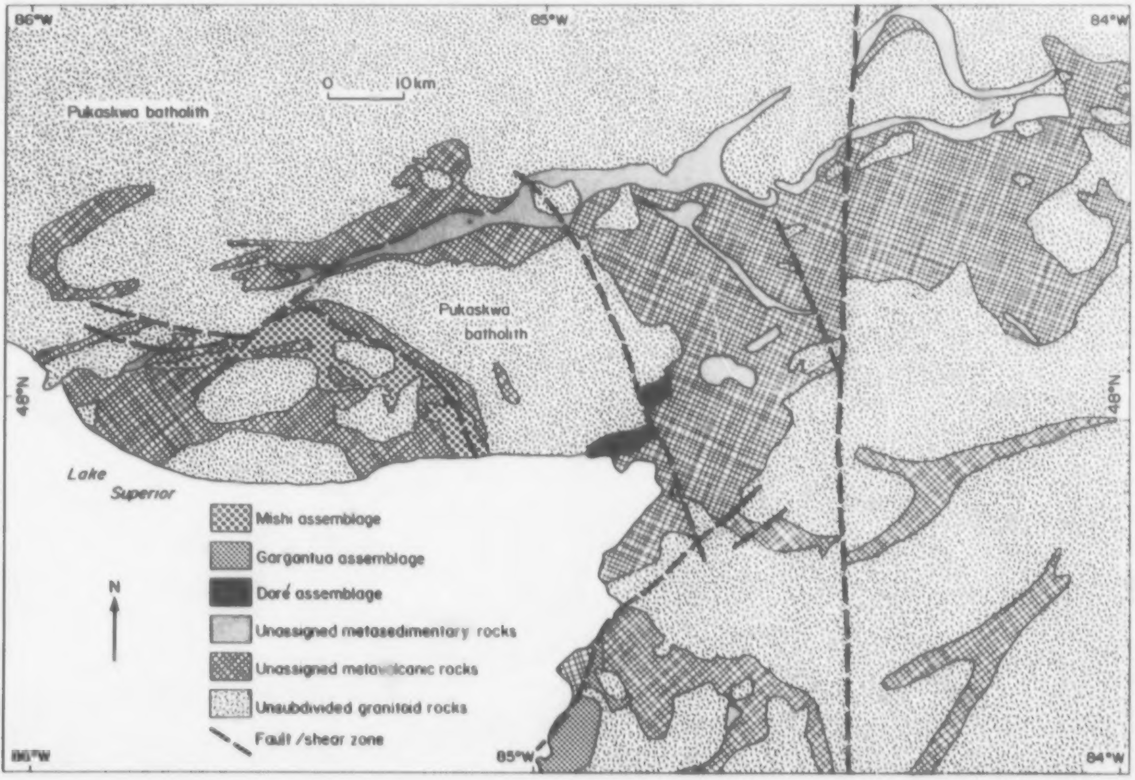


**Figure 25.4.** Tectonic assemblage map of the central Wawa Subprovince highlighting syn- to postaccretionary sedimentary sequences. Some newly defined assemblages appear on this figure and not on the Tectonic Assemblages of Ontario map. Adapted from the Tectonic Assemblages of Ontario map (*see* Map 2577, map case).

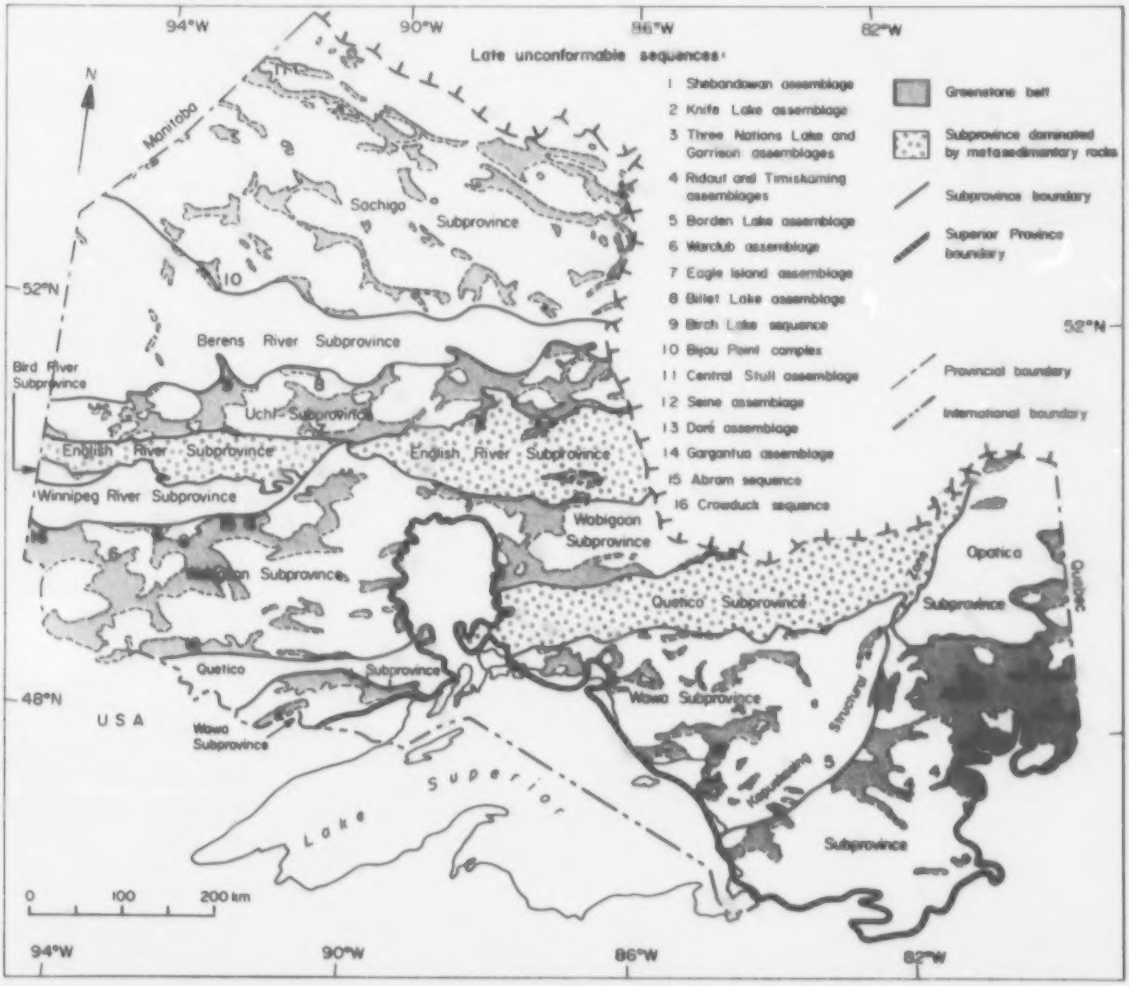
Regional patterns of stratigraphic facing in and between assemblages cannot be explained simply by folding; indeed, few large-scale early folds with preserved hinge zones have been delineated in the Superior Province. Whereas dominantly metasedimentary assemblages deform mainly by folding (e.g., Mishi and Gargantua assemblages, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume), those composed mainly of massive volcanic rocks deform internally along shear zones (e.g., Garden Lake belt, Wabigoon Subprovince, Thurston et al. 1987).

Assemblages do not deform in isolation but as units in a collage, as supported by correlation of shear sense in the fabrics of neighbouring assemblages (e.g., Tisdale, Krist and Porcupine assemblages, Abitibi Subprovince, Timmins area, *see* Jackson and Fyon, this volume; Piroshco and Kettles 1988). No evidence of pre-juxtaposition deformation has been documented in the Ontario part of the Superior Province.

There are 2 styles of deformation, operating either independently or contemporaneously, that may occur in an assemblage:

1. shear zones at assemblage margins
2. internal shear zones and folding concentrated in zones of low competency contrast (e.g., contacts of iron formation units with mafic volcanic or sedimentary rocks), or of high competency contrast (e.g., within wacke-siltstone units)

Many assemblage boundaries have been drawn at major shear zones. Other boundaries may initially have been unconformities which acted as loci for shearing during



94°W
90°W
86°W
82°W
N
Manitoba
Sochigo Subprovince
10
52°N
Berens River Subprovince
Bird River Subprovince
8
Uchi Subprovince
English River Subprovince
English River Subprovince
Winnipeg River Subprovince
6
Wabigoon Subprovince
Quetico Subprovince
Quetico Subprovince
Wawa Subprovince
Wawa Subprovince
USA
48°N
Lake Superior
Kapuskasing Structural Zone
5
Opatica Subprovince
Quebec
Subprovince
0 100 200 km
Late unconformable sequences:
1 Shebandowan assemblage
2 Knife Lake assemblage
3 Three Nations Lake and Garrison assemblages
4 Ridout and Timiskaming assemblages
5 Borden Lake assemblage
6 Warclub assemblage
7 Eagle Island assemblage
8 Billet Lake assemblage
9 Birch Lake sequence
10 Bijou Point complex
11 Central Stull assemblage
12 Seine assemblage
13 Doré assemblage
14 Gargantua assemblage
15 Abram sequence
16 Crowduck sequence
Greenstone belt
Subprovince dominated by metasedimentary rocks
Subprovince boundary
Superior Province boundary
Provincial boundary
International boundary

**Figure 25.5.** Geological map of the western part of the Superior Province, showing the major tectonic subdivisions of Card and Ciesielski (1986) and greenstone belts. Late unconformable sequences, associated unconformable, dominantly sedimentary sequences, and unassigned sedimentary sequences are identified. Some late unconformable sequences have been designated as lithotectonic assemblages, others not assigned to assemblages are designated as "sequences" or as "unassigned". Some assemblages are exaggerated in size for the purposes of clarity on this figure. Adapted from the Tectonic Assemblages of Ontario and Bedrock Geology of Ontario maps (*see* Maps 2575 to 2578, 2583, and 2541 to 2545, map case).

subsequent regional tectonism (e.g., boundary between Wawa and Catfish assemblages, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume).

## STYLES OF ASSEMBLAGE JUXTAPOSITION

Lacking evidence to the contrary, assemblages separated by shear zones are here provisionally considered to be mutually allochthonous, by which we infer only that juxtaposition was effected by tectonic processes; perhaps by simple thrust imbrication induced by a higher scale process, such as convection-driven plate motion leading to collision at subduction zones. We cannot yet differentiate shearing related to initial assemblage juxtaposition from that associated with later tectonism. Unfortunately, shear-sense indicators of displacement in these zones are generally absent or highly ambiguous (e.g., Hemlo–Heron Bay assemblage boundary, Hemlo greenstone belt, *see* Williams, Stott, Heather et al., this volume). Many shear zones show evidence of both dip-slip and strike-slip motions, which may be inferred to result from a composite of displacement histories and preferred strain partitioning. In rare cases, geochronological data permits a dip-slip shear zone to be inferred as a thrust fault; for example, combined structural analysis (Poulsen et al. 1980) and age determinations (Davis et al. 1989) indicates thrust faulting in the Rainy Lake area. More generally, cases can be found in most subprovinces where thrusting is needed to explain absolute age relations, cyclic repetition or other lithostructural features (e.g., Sachigo Subprovince: Ayres and Corfu 1991; Uchi Subprovince: Thurston and Breaks 1978, *see also* Stott and Corfu, this volume; Wabigoon Subprovince: Blackburn 1980, *see also* Blackburn et al., this volume; Abitibi Subprovince: Piroshco and Kettles 1988).

Some assemblages formed, or were preserved, as a result of normal and wrench faulting (e.g., Timiskaming and Shebandowan assemblages). Such assemblages are commonly in part unconformable on underlying greenstone belt rocks. A combination of sedimentological, geochemical, geochronological and structural data indicate that many of these assemblages may have developed via processes similar to those in pull-apart basins and small-scale rifts (*see* Table 25.2), and are essentially autochthonous.

## TIMING OF AGGREGATION

The aggregation of assemblages to form larger supracrustal collages generally preceded the emplacement of large volumes of tonalite-granodiorite plutons. Greenstone belts appear to be a remnant or artifact of plutonism; a collage of assemblages may encompass more than 1 remnant greenstone belt. For example, in the Uchi Subprovince (*see* Stott and Corfu, this volume), correlation of ages of volcanic assemblages between pluton-separated greenstone belts suggests that the same aggregations which constitute greenstone belts can be traced for hundreds of kilometres along the length of a subprovince. Some smaller assemblages, structurally overlying such long linear assemblage aggregates traceable from belt to belt, are inferred to have been tectonically interleaved with older rocks during or after assemblage aggregation (e.g., Boyer assemblage, Wabigoon Subprovince, *see* Blackburn et al., this volume). Calc-alkalic volcanic assemblages have commonly been intruded by consanguineous, penecontemporaneous plutons; an allochthonous calc-alkalic assemblage can carry with it its coeval plutons. However, the consanguineous plutons of an assemblage may intrude adjacent, older assemblages thereby rooting the assemblage as an autochthon (e.g., Woman and Confederation assemblages, Uchi Subprovince, *see* Stott and Corfu, this volume). In the few instances where plutons are older than adjacent supracrustal rocks, the latter unconformably overlie the former (e.g., Marmion Lake batholith and Steep Rock Group, *see* Blackburn et al., this volume).

## LARGER-SCALE AGGREGATION IN THE SUPERIOR PROVINCE

A discussion of geological, geochronological and geophysical constraints on the aggregation of assemblages into subprovinces and ultimately into the Superior Province follows. We propose that at a late stage in the development of granite-greenstone subprovinces, these interacted with adjacent sedimentary and pluton-dominated subprovinces. In this section we will also describe the formation of greenstone belts within the abundant granitoid masses of the granite-greenstone subprovinces and the inter-action of these subprovinces (e.g., Uchi, Wabigoon and Wawa subprovinces) with the predominantly sedimentary (e.g., English River and Quetico) and plutonic (e.g., Winnipeg River) subprovinces.

## DEVELOPMENT OF GREENSTONE BELTS

Thurston and Chivers (1990) documented the lithological diversity of greenstone belts, and recognized a variety of rocks formed in a range of different environments; this has been summarized above and discussed elsewhere in this volume. Shallow-water (continental) platforms, fore-arc prisms, rift and wrench fault-related sequences have all been identified using diagnostic rock types and sedimentary facies.

Despite this recognition of the environmental diversity of greenstone belts, we have only recently become aware of large-scale patterns of depositional environment. Volcanological (Johns 1985) and sedimentological (Fralick 1987) facies studies over large areas of subprovinces have only locally been attempted in detail. There may be more than 1 facies in an assemblage, a sign that further subdivision of presently defined assemblages is locally necessary.

# STRUCTURAL DEVELOPMENT OF GREENSTONE BELTS

## Structural Styles

In the introduction to this chapter, we noted the importance of worldwide structural analysis of greenstone belts. In the Superior Province, such analysis has revealed a structural development involving 3 contrasting styles:

1. early recumbent folding and "thin-skinned" thrusting, possibly in part related to assemblage juxtaposition
2. later upright folding, developed during and after tonalite batholith emplacement
3. strike-slip shear deformation, including folds and faults resulting from subsequent oblique convergence and interaction of subprovinces

### THRUSTS AND RECUMBENT FOLDS

Recumbent folding occurs at subprovince margins (e.g., Wabigoon–Quetico subprovince boundary, Rainy Lake area, Poulsen et al. 1980) and within greenstone successions in granite-greenstone subprovince cores (e.g., Michipicoten area, Arias and Helmstaedt 1990; Sioux Lookout–Dinorwic area, Chorlton 1991). Thrust fault repetition in the Beardmore–Geraldton Belt has been inferred from combined structural and sedimentological data (Devaney and Williams 1989); similar thrust juxtaposition of contrasting assemblages has been inferred from structural data from the Wabigoon–Quetico subprovince boundary in the Rainy Lake area (Davis et al. 1989). Thrust stacking of stratigraphic units is also suggested from geochronological data in several places (e.g., Confederation assemblage, Uchi Subprovince, Noble et al. 1989, *see* Stott and Corfu, this volume; Favourable Lake greenstone belt, Sachigo Subprovince, Corfu and Ayres 1991) and also suspected along the Winnipeg River–Wabigoon subprovince boundary, in the Sioux Lookout area (Beakhouse 1988, this volume; Davis et al. 1988; *see* Blackburn et al., this volume).

Many assemblage boundaries have been placed at steeply dipping shear zones which may have initiated as thrust faults. In most shear zones, however, the early shear-sense is not apparent This is mainly due to the overprinting effects of later deformation and, in part, our inability to determine not only the slip-sense in highly schistose rocks, but also the occurrence of thrusting in rocks that are now steeply dipping. An example of thrust-induced repetition has been documented from the Michipicoten greenstone belt, Wawa Subprovince. There, stratigraphically and geochronologically equivalent rocks of the Catfish assemblage are repeated within the eastern part of the McCormick synform; the 2 equivalents are known there as the McCormick and Michipicoten segments (*see* Williams, Stott, Heather et al., this volume).

The recognition of early recumbent fabric elements is dependent on the ability to recognize the full effects of their rotation into a subvertical orientation by later upright folding or doming by granitoid batholiths (Chorlton 1991). The recognition of thrust faults within an assemblage or affecting several assemblages is not proof of thrust juxtaposition of assemblages. However, thrust faults are commonly indicative of large-scale horizontal transport and shortening, which elsewhere produces assemblage juxtaposition (Howell et al. 1985).

### UPRIGHT FOLDING

Upright folding around steeply to shallowly plunging axes (Wabigoon Subprovince, *see* Blackburn et al. this volume; Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume) was responsible for steepening the structural attitudes of lithological units and early, originally horizontal structural fabrics in greenstone belts into a near vertical orientation. The relative timing of batholith emplacement and upright folding varies across the Superior Province. Folds developed: 1) before batholiths of tonalite-granodiorite were emplaced, and are therefore cut by the latter (e.g., Dryberry Batholith, Sanborn-Barrie 1991a); 2) during batholith emplacement, in which case folds bear spatial and trend relationships with batholith contacts (e.g., Dryberry Batholith, Sanborn-Barrie 1991a; cf., Williams and Breaks 1990; de Wit et al. 1987); or 3) after batholith emplacement, in which case batholith and host were folded together (e.g., Rainy Lake batholithic complex, Smith 1990, Schwerdtner 1990; Kakagi Lake area, Davis and Edwards 1985).

### LATE-STAGE SHORTENING

Upright folding was probably in part penecontemporaneous with, and preceded the development of, large-scale dextral shear zones (e.g., eastern Lake of the Woods area, Sanborn-Barrie 1991b). The concentration of shear zones, which in part defines subprovince boundaries, (e.g., Wabigoon fault, Quetico Fault), may also be traced within granite-greenstone subprovinces along the axial parts of large, curvilinear greenstone belts (e.g., Manitou Straits fault, Pipestone–Cameron deformation zone; *see also* shear zones along the northern edge of the Mishi assemblage, Wawa Subprovince, Figure 25.4). Some shear zones may include reactivated early faults that form assemblage boundaries. The combined function of the shear zones and upright folds was to accommodate horizontal regional stress directed obliquely to the subprovince boundary. The presence of dextral shear zones throughout the Superior Province, in addition to the ubiquitous upright folding, have been identified as signatures of a transpressional tectonic regime (Stott et al. 1987) and several workers, namely, Hudleston et al. (1988), Tabor and Hudleston (1991), and Borradaile and Spark (1991) have demonstrated this in detail.

### AGE RANGES OF DEFORMATION

Much of the deformation in the Superior Province is the result of phases of the Kenoran Orogeny (Stott and Corfu 1988), but in detail, multiple episodes of volcanism, deformation and plutonism have been recorded even within individual subprovinces. This is illustrated in the Uchi Subprovince by Corfu and Stott (1989) and Stott and Corfu (this volume). For example, in the Pickle Lake greenstone belt, regional deformation and greenschist metamorphism have locally been overprinted by plutonism dated at 2740 Ma, which predates volcanism of the St. Joseph assemblage (2725 to 2711 Ma) and the subsequent deformation and intrusion by plutons along the southern flank of the subprovince (*see* Figure 25.3). Another example comes from the central Sachigo Subprovince where the crescent-shaped, posttectonic North Caribou batholith, dated at 2870 Ma (D. Davis, Royal Ontario Museum, personal communication, 1989), and intruded between a tonalitic gneiss and the arcuate North Caribou greenstone belt, has imposed a biotite-amphibolite hornfels aureole upon an existing regional metamorphic gradient in the belt (Stott et al. 1989; Breaks et al., in prep.). The large extent of regional metamorphism, subsequent intrusion and thermal overprinting supports an earlier (pre–Kenoran) orogeny that terminated at about 2870 to 2850 Ma.

Elsewhere, geochronological control on the absolute timing of deformation has been sought along the Wawa–Quetico (Corfu and Stott 1986), Wabigoon–Quetico (Davis et al. 1989) and English River–Uchi (Stott et al. 1987) subprovince boundaries (*see* Figure 25.1). As well, the maximum age of transpressional deformation may be constrained by the age of supracrustal rocks spatially associated with the wrench faults or overlying them (e.g., Timiskaming assemblage, Abitibi Subprovince, *see* Jackson and Fyon, this volume). Initially, transpression may not have led to the preservation of supracrustal rocks, but those that have been preserved give only the minimum age for the initiation of transpression; the 2689 million-year-old Shebandowan assemblage in the Wawa Subprovince has been deformed by transpressive deformation (*see* Williams, Stott, Heather et al., this volume).

Most greenstone belts contain assemblages for which a minimum age is known; these have been affected by folding and shearing which in some cases has been postdated by the emplacement of undeformed, discordant granitoid plutons. Thus, deformation is commonly physically and temporally constrained between assemblage formation and the intrusion of plutons. In some cases, this is a tight control, on the order of 10 to 20 million years (*see* Corfu and Andrews 1987); more commonly, bracketing is less precise, for example, the time gap between the 2830 million-year-old volcanism and the 2709 million-year-old assemblage juxtaposition of the Munro Lake and Oxford Lake–Stull Lake terranes (*see* Thurston et al., this volume). The time span in which deformation took place is not dissimilar to that documented from modern orogens, such as the Taconic phase of the Appalachian Orogeny (Williams and Hatcher 1982).

Most of the deformation in greenstone belts of the southern Superior Province appears to have occurred after juxtaposition of young assemblages with older assemblage aggregations; that is, at about 2700 Ma, which is prior to the extensive emplacement of granitoid rocks at about 2690 Ma. However, field and geochronological relationships described from the Sachigo Subprovince and from greenstone belts farther south, in the Wawa and Wabigoon subprovinces, indicate a progressive southward- and northward-younging diachroneity of deformation in the Superior Province (Stott et al. 1987).

## RELATIONSHIP OF DEFORMATION TO METAMORPHISM

Only in the Abitibi Subprovince is there evidence of an early burial metamorphism (Jolly 1978). Elsewhere, metamorphic pressures (Figure 25.6) developed prior to the emplacement of granitoids may reach up to 0.6 GPa (e.g., Hemlo greenstone belt, Pan 1990). The extent to which the higher grade greenstone belts and the voluminous tonalite batholiths were subjected to pressures of this order has not yet been established; in the Berens Lake area, D. Stone (*see* Thurston et al., this volume) determined metamorphic pressures of 0.4 to 0.6 GPa. On the scale of many of the larger individual greenstone belts, most, if not all of the ductile deformation occurred during relatively low-grade metamorphic conditions (e.g., Abitibi Subprovince, *see* Jackson and Fyon, this volume; *see also* Jolly 1980).

Evidence for 2 events of metamorphism has been cited in a previous section: a hornfelsic aureole around the North Caribou batholith overprints an existing metamorphic signature (Stott et al. 1989). In the Uchi, Wabigoon, Wawa and Abitibi subprovinces, contact metamorphic aureoles locally overprinted earlier syntectonic metamorphic signatures , which underwent static annealing during posttectonic cooling (Ayres 1978; Jolly 1978, 1980; Thurston and Breaks 1978). In the Uchi Subprovince, the thermal aureole of the Ochig Lake pluton (2740 Ma) has been superimposed on the assemblages of the Pickle Lake greenstone belt (2890 Ma; *see* Stott and Corfu, this volume).

In the smaller greenstone belts engulfed by a greater proportion of tonalite-granodiorite plutons, metamorphic temperatures and pressures were commonly much higher, usually corresponding to amphibolite to granulite facies (e.g., Manitouwadge–Hornepayne greenstone belt, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume; Y. Pan, University of Western Ontario, personal communication, 1991). We infer therefore that greenstone belt size is in part related to depth of burial during metamorphism.

## DEVELOPMENT OF SUBPROVINCES

Early workers subdivided the Superior Province into subprovinces, which were distinguished largely on the basis that greenstone belts are more common in elongate zones some hundred kilometres wide and a thousand kilometres long (*see* Card and Ciesielski 1986). These zones are herein known as granite-greenstone subprovinces (e.g., Uchi, Wabigoon, Abitibi and Wawa subprovinces). They are separated by zones dominated either by metasedimentary rocks (e.g., English River and Quetico subprovinces), or by plutonic igneous rocks and their high-grade metamorphic equivalents (e.g., Winnipeg River subprovince). These too, have been credited with subprovince status.

Early workers also recognized that adjacent subprovinces were separated by major topographic lineaments, which were inferred to be faults with strike-slip displacements commonly on the order of tens to hundreds of kilometres. Motions on these faults have been the objects of paleomagnetic study (Dunlop 1984), but only recently have rotations between adjacent fault-bounded blocks within subprovinces been determined (Hale and Lloyd 1990).

Granite-greenstone subprovinces contain granitoid intrusive rocks and diverse supracrustal rocks in proportions of about 3:1; their geology is relatively well known. In contrast, high-grade metasedimentary and pluton-dominated subprovinces are not as well known, contain units with fewer marker horizons for structural studies and are considered by many to have a lower potential for exploitable mineral deposits. Indeed, large areas of these subprovinces and their boundaries have only recently been the subject of sedimentological, structural, igneous and metamorphic study (e.g., Quetico Subprovince; *see* Percival 1989; *see also* Williams, this volume).

## Granite-Greenstone Subprovinces

The following brief review of stratigraphic, geochronological and structural data indicates that most greenstone belts developed from a series of tectonically juxtaposed assemblages, which were subsequently intruded and largely deformed by granitoid batholiths, for example, the northwestern part of the Superior Province (*see* Thurston et al., this volume) and the Abitibi Subprovince (*see* Jackson and Fyon, this volume).

### STRATIGRAPHY OF GREENSTONE BELTS

The supracrustal tectono-stratigraphy of greenstone belts consists predominantly of distinct assemblages of diverse, ultramafic to mafic, calc-alkalic andesitic to rhyolitic volcanic rocks, and subordinate wackes and conglomerates.

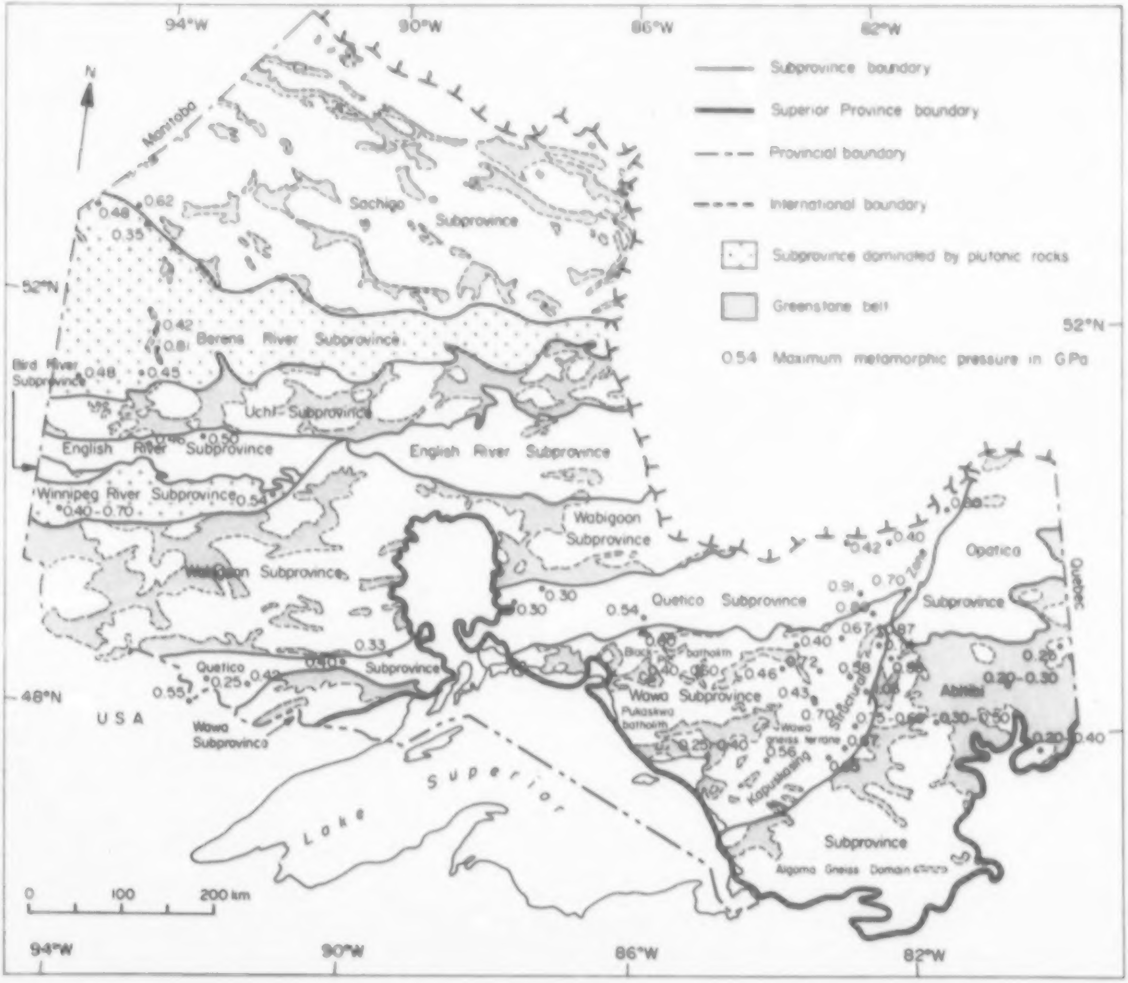

94°W
90°W
86°W
82°W
N
Manitoba
Subprovince boundary
Superior Province boundary
Provincial boundary
International boundary
Subprovince dominated by plutonic rocks
Greenstone belt
0.54 Maximum metamorphic pressure in GPa
52°N
48°N
Sachigo Subprovince
Berens River Subprovince
Bird River Subprovince
Uchi Subprovince
English River Subprovince
Winnipeg River Subprovince
Wabigoon Subprovince
Quetico Subprovince
Wawa Subprovince
Opatica Subprovince
Abitibi
Quebec
Black-Pic batholith
Pukaskwa batholith
Wawa gneiss terrane
Kapuskasing
Structural Zone
Algoma Gneiss Domain
Lake Superior
USA
0
100
200 km
0.48
0.62
0.35
0.42
0.81
0.48
0.45
0.46
0.50
0.54
0.40-0.70
0.30
0.30
0.33
0.40
0.42
0.25
0.55
0.54
0.42
0.40
0.91
0.70
0.67
0.87
0.40
0.72
0.46
0.43
0.70
0.56
0.25-0.40
0.28
0.20-0.30
0.30-0.50
0.20
0.40

**Figure 25.6.** Geological map of the western part of the Superior Province, showing major tectonic subdivisions of Card and Ciesielski (1986) and greenstone belts. Plutonic rock-dominated subprovinces, and pluton-rich areas within greenstone-granite subprovinces are identified. Included are selected determinations of maximum metamorphic pressure (sources cited in text).

Supracrustal rocks in greenstone belts consist of a collage of tectonic assemblages, generally of less than 5 km true thickness, consisting of mafic volcanic sequences, sporadic mafic to felsic central volcanic complexes, and associated aprons of volcanogenic sediment (Blackburn et al. 1985). These rocks have been assigned to stratigraphic units of different age. Intrusive rocks surrounding greenstone belts commonly consist of complex multiphase tonalite-granodiorite and granodiorite-granite batholiths.

## STRUCTURAL RELATIONSHIPS WITHIN GREENSTONE BELTS

The boundaries of assemblages within greenstone belts are commonly sheared, unconformable, or intrusive. Greenstone belts consisting of volcanic and sedimentary rocks of contrasting ages may still contain relatively undisturbed, though deformed, stratigraphic sequences (e.g., Wawa assemblage unconformably overlain by the Catfish assemblage, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume). Stratigraphic facing directions within greenstone belts may exhibit "homofacing", that is, areas exhibiting uniform younging directions, back-to-back, or front-to-front geometries that were tectonically produced. Many greenstone assemblages have been mutally juxtaposed back-to-back (e.g., Balmer and Confederation assemblages at Red Lake, Uchi Subprovince, *see* Stott and Corfu, this volume) and front-to-front (e.g., Miron and Pukaskwa–Point Isacor segments, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume), separated by shear zones. Some assemblages are separated by unconformities (e.g., Greenwater and Shebandowan assemblages, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume), which later became the loci for shearing (e.g., Setting Net and Northwind assemblages, Favourable Lake greenstone belt, *see* Thurston et al., this volume).

## GREENSTONE ASSEMBLAGE AGE DISTRIBUTION WITHIN SUBPROVINCES

There is increasing evidence to support the hypothesis that within each granite-greenstone subprovince in the Superior Province, ages of greenstone assemblages generally decrease southward (*see* Figure 25.3). A poorly constrained southward younging of assemblages is seen within the Wabigoon Subprovince. There, old assemblages, such as the Fourbay assemblage, are bordered by progressively younger ones to the south (*see* Blackburn et al. this volume). In addition, the triangular shape and basal tectonic boundaries of such units as the Boyer and Kakagi assemblages imply that these assemblages may be allochthonous, having been structurally emplaced from the south over older assemblages.

Similarly in the Wawa Subprovince, the oldest assemblages, such as the Burchell and Hemlo assemblages, occur along the northern boundary, whereas the youngest volcanic rocks occur farther south in the Gamitagama greenstone belt (*see* Williams, Stott, Heather et al., this volume).

Perhaps the clearest example of this feature comes from the Uchi Subprovince (*see* Stott and Corfu, this volume) and the northwestern Superior Subprovince (*see* Thurston et al., this volume). Figure 25.3 shows volcanic assemblages decreasing in age southward, from 2990 Ma in the Red Lake greenstone belt (Balmer assemblage) to 2713 Ma in the Lake St. Joseph greenstone belt (St. Joseph assemblage), near the southern flank of the Uchi Subprovince. This southward younging of episodic volcanism is prominent across the central Uchi Subprovince, particularly south from Pickle Lake (*see* Figure 6.3, this volume).

Some of the oldest rocks in Ontario are felsic plutons approximately 3.0 Ga in age, found in the central part of the Wabigoon Subprovince (e.g., Marmion Lake batholith, Wabigoon Subprovince, *see* Blackburn et al., this volume). A continental platform phase of Mesoarchean greenstone supracrustal rocks has been locally preserved on these old rocks in this region (Thurston and Chivers 1990). Recent analyses by D. Davis (Royal Ontario Museum, personal communication, 1991) have shown that quartz-rich arenites and conglomerates (Jutten group) in the Savant Lake greenstone belt contain zircon grains as old as 3.3 Ga.

## GRANITOID REGIONS

There are 2 main types of granitoid rocks, totalling about 75% by area, that can be distinguished in granite-greenstone subprovinces: 1) syndepositional to syntectonic, complex, composite foliated to gneissic batholiths of tonalite and granodiorite; and 2) syntectonic to late tectonic plutons and batholiths of foliated to massive granodiorite to granite. The earlier plutonic suite is magmatically and temporally related to the late stages of volcanism within the greenstone belts they engulf (Beakhouse et al. 1989; Edwards 1985).

The relative proportions of the 2 compositional types varies both within and between subprovinces and may be a function of erosion level (H.R. Williams 1991); however, the earlier suite tends to surround and therefore separate greenstone belts. Examples of regional-scale tonalitic batholiths are: Black–Pic batholith in the Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume), the Dryberry Batholith (Sanborn-Barrie 1991a) and the Rainy Lake batholithic complex (Schwerdtner 1990) in the Wabigoon Subprovince. The later, granitic (*sensu stricto*) suite is subordinate in volume to the earlier emplaced suite, and constitutes discrete plutons within greenstone belts, both as late-stage crescentic bodies along tonalite-greenstone contacts (Schwerdtner et al. 1983) and pancake-shaped batholiths that were magmatically "stoped" into place (e.g., White Otter Lake batholith, *see* Blackburn et al., this volume; Lount Lake batholith, *see* Beakhouse, this volume, *see also* Beakhouse and McNutt 1991; and the Williams suite, *see* Stott and Corfu, this volume).

The granitic basement of supracrustal assemblages has rarely been observed in the Superior Province; the Marmion Lake batholith, which underlies the Lumby assemblage, is a notable example (*see* Blackburn et al., this volume). Instead, many tonalite-granodiorite batholiths are large antiformal masses possessing margins that are structurally concordant with the supracrustal rocks. For example, in the Wawa

Subprovince, the Black–Pic batholith (*see* Williams, Stott, Heather et al., this volume) separates the Hemlo greenstone belt from high-grade tonalite gneisses at Manitouwadge which contain concordant supracrustal units (*see* Williams, Stott, Heather et al., this volume). The Black–Pic batholith consists of a dome of gneissic to foliated rocks to which the neighbouring supracrustal rocks appear to have a complementary synformal relationship.

Some tonalite-granodiorite masses are folded together with supracrustal rocks (Williams and Breaks 1990) and share similar subparallel patterns of planar and linear fabrics (e.g., Uchi Subprovince, *see* Stott and Corfu, this volume; Wawa Subprovince, Heather 1989).

## AGE RELATIONSHIPS BETWEEN GRANITOIDS AND GREENSTONE ASSEMBLAGES

Some greenstone assemblages are spatially and temporally associated with masses of tonalite-granodiorite; most tonalitic batholiths postdate, or are synchronous with greenstone assemblages. Geochemical (Edwards 1985) and isotopic (Davis and Edwards 1986) evidence permits the hypothesis that synvolcanic tonalitic to granodioritic plutons in the Wabigoon Subprovince may be the root zones of volcanoes (cf., Hildebrand 1984). Although spatial and temporal relationships between batholiths and volcanic centres (cf., Ayres 1977) have yet to be analyzed in detail, intrusion of synvolcanic plutons occurs preferentially in the most proximal part of the volcanic system, engulfing vent facies volcanic rocks. The consequence of this hypothesis is that distal parts of volcanic units flanking the source plutons might be preferentially preserved.

Some tonalitic batholiths are clearly older than the supracrustal sequences which unconformably overlie them (e.g., Marmion Lake batholith, Wabigoon Subprovince, Jolliffe 1955; *see also* Davis and Jackson 1988). Other evidence of a much older tonalite-granodiorite suite is exhibited by some members of the granite-granodiorite suite that contain inherited zircon grains; some granitic rocks were derived from the remelting of a felsic precursor (Beakhouse and McNutt 1991; Corfu and Wood 1986).

## TECTONIC SIGNIFICANCE OF GRANITOID ROCKS

The Bedrock Geology of Ontario map (Maps 2541 to 2545, map case) reveals a wide range in ratio of granitoid rocks to greenstone belt rocks in granite-greenstone subprovinces. An example of a high greenstone/granitoid ratio is the central part of the Abitibi Subprovince in eastern Ontario; in contrast, an example of a low ratio is the gneissic terrane around Manitouwadge (Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume) or northeast of Atikokan (Wabigoon Subprovince, *see* Blackburn et al., this volume).

The emplacement of large volumes of tonalitic magma into the crust beneath and/or within the supracrustal rocks must have represented a regional-scale thermal source from which energy passed into the greenstone belts, causing or adding to metamorphism caused by tectonic thickening processes.

Thin, sinuous (1 to 2 km wide) greenstone belts, lying within regions composed predominantly of tonalitic batholiths, are commonly of medium to high pressure (0.4 to 0.6 GPa) amphibolite facies (e.g., Manitouwadge–Hornepayne and Hemlo greenstone belts, *see* Williams, Stott, Heather et al., this volume). They occur as concordant and commonly shallowly dipping layers and enclaves within and surrounding complex domal plutons. Single-stage (1 assemblage) stratigraphic units and vestigial greenstone units can be traced as inclusion zones across batholiths, and contain few late discordant plutons (e.g., Manitouwadge–Hornepayne greenstone belt, Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume).

Large greenstone belts, up to 100 km across, were metamorphosed at relatively low pressures (0.25 to 0.4 GPa) to greenschist facies (e.g., Abitibi Subprovince, *see* Jackson and Fyon, this volume), and were intruded by relatively small amounts of granitic rock, despite being surrounded by granitoid rocks (H.R. Williams 1991). The large, complex aggregations of relatively low-pressure-metamorphosed greenstone assemblages (Feng and Kerrich 1990) commonly exhibit upright folding and block faulting punctuated by late discordant plutons (e.g., Abitibi greenstone belt, *see* Jackson and Fyon, this volume).

High metamorphic pressures determined from amphibolite to granulite facies rocks in greenstone belts indicate that these rocks represent deeply eroded crust exhumed after crustal thickening and consequent erosion. The distribution of low and high grades of metamorphism in the Archean may be related to large, fault-related uplifts of the type suggested by Percival and Card (1983). However, we suggest that the distribution of metamorphic grades is in part due to regional variation in the extent and timing of tectonic and magmatic thickening. Thin, highly metamorphosed greenstone belts need not simply represent deeper tectonic levels of large, contiguous, complex and less metamorphosed greenstone belts (H.R. Williams 1991; cf., de Wit et al. 1987) but may instead be manifestations of arrested supracrustal development.

## GRANITE-GREENSTONE SUBPROVINCE BOUNDARIES

The boundary zones of granite-greenstone subprovinces display features not seen within their interiors; for example, structural styles within the interior part of the Wabigoon Subprovince (*see* Blackburn et al., this volume) contrast with those displayed at its boundaries. Boundary zones exhibit features characteristic of late-stage transpressional shortening in a strike-slip regime (Hudleston et al. 1988).

In the Beardmore–Geraldton Belt, for example, shortening across the Wabigoon–Quetico subprovince boundary is postulated to have been accommodated by the production of "thin-skinned", south-verging thrust faults that repeat stratigraphic units. These units and the thrust faults were rotated into a near vertical orientation and then subjected to ductile, dextral transcurrent shear (*see* Figure 10.4, this volume; Williams 1986; Devaney and Williams 1989). Similar structural relationships occur in rocks in the Rainy Lake (Poulsen et al. 1980) and Kashabowie (Borradaile and Spark 1991) areas and are suspected in the Sioux Lookout region (Davis et al. 1988); modern analogues include the Setogawa and Mikasa groups of the Shimanto Belt in southwest Japan (Osozawa 1988).

Structural patterns in greenstone belts may be a function of: 1) edge effects resulting from large-scale, terrane or superterrane accretion; and 2) shortening induced by lateral expansion of plutons. We cannot on a regional scale reliably distinguish the relative significance of these 2 processes, which together have resulted in complex inhomogeneous strain patterns. Relatively late-stage compressional features, such as folds and faults at subprovince boundaries, may be due to subprovince interaction; some may be due to local shortening resulting from large-scale emplacement of plutons within the central parts of both the granite-greenstone and sediment-dominated subprovinces (cf., Schwerdtner 1976).

Subprovince boundaries are commonly the locus of high metamorphic gradients or sharp contrasts in metamorphic grade. Metamorphic gradients both along and across subprovince boundaries vary: from greenschist to amphibolite facies across the Uchi–English River (Thurston and Breaks 1978; *see* Breaks, this volume) and Wabigoon–Quetico subprovince boundaries (Pirie and Mackasey 1978); and from amphibolite to granulite facies across the Wawa–Quetico (*see* Williams, Stott, Heather et al., this volume; Williams, this volume) and Winnipeg River–English River subprovince boundaries (*see* Breaks, this volume).

Plutonic rocks spatially associated with subprovince boundaries, such as along the Wabigoon–Quetico subprovince boundary (e.g., Croll Lake stock, Williams 1990) are commonly slightly elliptical, exhibiting distorted teardrop-shaped outlines. These shapes, which are rarer in the central parts of granite-greenstone subprovinces (e.g., Cedar Lake pluton, Wawa Subprovince, Muir 1982), are here interpreted to have resulted from magma emplacement during ductile deformation. The asymmetry of distortion from an originally equidimensional shape is a guide to the sense of displacement on the boundary during transpression. Despite emplacement in an actively deforming regime, not all of the magmatic bodies are internally deformed, and may exhibit massive central parts. For example, along part of the Quetico–Wawa subprovince boundary near Shebandowan (*see* Williams, Stott, Heather et al., this volume), plutons of a late-stage diorite-monzogranodiorite suite were emplaced during transpression, but were not apparently deformed.

Shoshonitic volcanism and plutonism of Neoarchean age has been recognized in 2 major regions: 1) in the northern part of the Wawa Subprovince (Hodgkinson 1968; Shegelski 1980; *see* Williams, Stott, Heather et al., this volume); and 2) in the Abitibi greenstone belt, associated with the major east-trending Larder–Cadillac deformation zone (Basu et al. 1984; *see* Jackson and Fyon, this volume). Shoshonitic volcanic rocks also occur in the Sachigo Subprovince (Oxford Lake Group; Hubregtse 1976) and at North Spirit Lake (Smith and Longstaffe 1974). Rocks of this composition are considered by some to be signatures of vestigial island arc magmatism during arc closure (Wyman and Kerrich 1989), and by others (Thurston and Chivers 1990) to be related to intracontinental wrench faults.

The amalgamation of subprovinces and subsequent deformation at subprovince boundaries is constrained by ages of crosscutting plutons that predate and postdate deformation. A pluton postdating deformation at the Wawa–Quetico subprovince boundary in northwestern Ontario (Corfu and Stott 1986) was subjected to structural and geochronological analysis. The combination of structural and geochronological data on plutons and volcanic rocks delimited the timing of boundary formation. Recent work has shown that some of these boundaries, such as the Quetico and Sydney Lake–Lake St. Joseph faults, remained active, at least into the Mesoproterozoic (Kamineni et al. 1990).

## Metasedimentary Subprovinces

Subprovinces dominated by sedimentary rocks (*see* Figure 25.5) have received little attention until recently (Percival 1989a) because of their relatively low economic potential. The recognition of thrust fault-bounded sequences at the margins of such subprovinces yielded clues to their origin as a stack of thrust-imbricated sedimentary rocks (Williams 1986, 1990; Devaney and Williams 1989; Percival and Williams 1989).

Sedimentary rock-dominated subprovinces include rapidly deposited detritus derived from adjacent granite-greenstone terranes. During sedimentation, tectonically or diapirically emplaced and eroded ultramafic rocks produced local accumulations of serpentinous sediment (*see* Williams, this volume; cf., Ernst and Jahn 1987).

Intrusions within the sedimentary rock-dominated subprovinces consist of early, mantle-derived tonalitic rocks (e.g., Chamberlain Narrows and Bluffy Lake batholiths, English River Subprovince, *see* Breaks, this volume). These were followed by suites of biotite and "two-mica" granites (e.g., Sturgeon Lake batholith, Percival 1989a; *see also* Williams, this volume), which are genetically and temporally related to the high-grade metamorphism and melting of supracrustal rocks. Regionally, as in the western part of the Quetico Subprovince, granitoid rocks may dominate over sedimentary rocks (*see* Figure 10.2a, this volume).

A combination of crustal thickening by thrust faulting and folding may have induced melting of lower parts of these subprovinces, transferring heat towards the surface as "S-type" granitic magma, around which metamorphic grade is locally increased (e.g., Quetico, *see* Williams, this volume; English River, *see* Breaks, this volume).

## ENVIRONMENT OF FORMATION

Sedimentological and stratigraphic evidence of rapid sediment deposition in the Quetico Subprovince was presented by Williams (this volume). Sediment was deposited from south-flowing grain and turbidity flows as submarine fans and on the margins of interdistributary channels (Devaney and Williams 1989). These sedimentary rocks contain detrital zircon grains as old as 3.0 Ga (Davis et al. 1990), whereas sediment deposition must predate a tonalite intrusion dated at 2688 Ma; a lower age limit of sedimentation is undetermined, but is probably in the region of 2710 Ma and perhaps as young as 2698 Ma (Davis et al. 1990; Percival and Williams 1989). Similar age ranges and absolute values have been interpreted from data acquired from the English River Subprovince (*see* Breaks, this volume) and from the Pontiac Subprovince of Quebec (Gariépy et al. 1984).

### DEFORMATION, MAGMATISM AND METAMORPHISM

The large volumes of sedimentary rock in the sedimentary rock-dominated subprovinces underwent deformation, magmatism and metamorphism within a short time period, starting during the late stages of sedimentation, and continuing until the cooling metamorphic welt could no longer deform in a ductile fashion.

Synsedimentary deformation ($D_1$) is recorded as slump deposits and as south-verging mesoscopic to regional-scale folds (Southwick 1991). Further deformation ($D_2$) resulted in the formation of: 1) planar fabrics that commonly transpose bedding; 2) steeply and variably plunging isoclinal folds of transposed bedding; and 3) migmatitic veining during folding. Belt-parallel upright folding ($D_3$) displays a Z-style asymmetry and is typical of the high-grade southern parts of these metasedimentary subprovinces.

The intrusive rocks within these subprovinces have been predominantly deformed by both the $D_2$ and $D_3$ deformations. The degree of metamorphism observed in these rocks varies, ranging from greenschist facies to upper amphibolite facies and locally, granulite facies. The distribution of metamorphic grade is asymmetric, with the highest grades dominating in the southern parts of the subprovinces (Williams 1990).

### ORIGIN OF METASEDIMENTARY SUBPROVINCES

A combination of sedimentological, structural, metamorphic and geochronological evidence invites comparison of the Quetico, English River and Pontiac subprovinces with modern fore-arc accretionary prisms in island arc environments (*see* Breaks, this volume; Percival and Williams 1989; Williams 1990, this volume). The similarity between modern and Archean volcanic arcs (Table 25.3) in terms of scale and tectonic relationships with adjacent sedimentary belts, supports an accretionary origin for the Archean metasedimentary subprovinces.

### RELATIVE CHRONOLOGY OF METASEDIMENTARY AND GRANITE-GREENSTONE SUBPROVINCES

The study of the timing of structural and metamorphic events within these 2 contrasting types of subprovince is hampered by the strong deformation that masks the original relationships at their margins. Geochronological data (Davis et al. 1990) and correlation of fabric development across subprovince margins (e.g., Sanborn-Barrie 1988) support a broad contemporaneity of large-scale structural and metamorphic development in greenstone belts with that in the sedimentary subprovinces (i.e., after 2.7 Ga, the Kenoran Orogeny). Field relationships suggesting a pre–Kenoran orogeny within the Uchi Subprovince (*see* Stott and Corfu, this volume) and farther north, indicate that those granite-greenstone subprovinces enjoyed at least 2, perhaps more, tectono-thermal events. After 2.7 Ga, both sedimentary and granite-greenstone subprovinces underwent shortening, thickening, shearing and cooling.

## Granitoid-Rich, Plutonic Subprovinces

A third type of subprovince is characterized by highly metamorphosed and variably deformed felsic plutonism (*see* Figure 25.6), for example, the Winnipeg River Subprovince (*see* Beakhouse, this volume) and the region formerly known as the Berens River Subprovince (Card and Ciesielski 1986); the latter has been described by Thurston et al. (this volume). In Ontario, granitoid-rich plutonic subprovinces are smaller than other types of subprovince and are comparable in scale to pluton-rich areas of the Abitibi, Wabigoon and Wawa subprovinces.

The Winnipeg River Subprovince contains igneous rocks as old as 3172 Ma (Corfu 1988), but most of the plutonism (Beakhouse et al. 1988; Beakhouse and McNutt 1991) occurred from 2710 to 2660 Ma (Corfu 1988). Supracrustal rocks are uncommon, but boundaries with adjacent subprovinces are zones of high strain coincident with thin supracrustal sequences (e.g., Bird River Subprovince, *see* Beakhouse, this volume).

Regional variation in metamorphic grade is characteristic of these subprovinces as enumerated below. Along the length of the Winnipeg River Subprovince, for example, metamorphic grade varies from granulite to epidote-amphibolite facies from east to west (Beakhouse et al. 1983). Similarly, in the area formerly known as the Berens River Subprovince (*see* Thurston et al., this volume) there are direct and indirect indications of both metamorphic temperature and pressure variation (*see* Figure 25.6; D. Stone, Ontario Geological Survey, personal communication, 1990).

### COMPARISON WITH GRANITOID AREAS IN GRANITE-GREENSTONE SUBPROVINCES

Granitoid-rich regions within granite-greenstone subprovinces are geologically and physically comparable with the Winnipeg River Subprovince; examples include the central Wabigoon region, the Pukaskwa and Black–Pic batholiths in the Wawa Subprovince (*see* Figure 25.6) and the Chapleau Gneiss Domain . There are no unequivocal lithological, structural or metamorphic criteria distinguishing plutonic masses within granite-greenstone subprovinces from those forming entire subprovinces.

**Table 25.3.** Parallels between Archean and modern arc complexes and related environments.

| Criterion | Archean Volcanic Arc (prior to and including subprovince aggregation) | Modern Volcanic Arc (e.g., Japan and southwest Pacific) |
|---|---|---|
| Scale | Build progressively until late Neoarchean, up to 200 km wide and 2000 km long | Discontinuous arcs up to 10 000 km long and between 50 to 300 km wide, such as Japan (Taira et al. 1988) and southwest Pacific (Hamilton 1988; Macfarlane et al. 1988; Honza 1991) |
| Rock types and their distribution | **Basement:** rare Mesoarchean tonalitic to granitic substrate; oldest volcanic assemblages tend to be on northern side of subprovinces | Precambrian to pre–Jurassic terranes at rear of arc: Hida and Sangun terranes of Japan (Saito 1985; Taira 1985; Taira et al. 1983) |
| | **Volcanic:** stages of early primitive komatiitic to tholeiitic subaqueous, commonly overlain by subordinate, unconformable or allochthonous diverse, progressively emergent, calc-alkalic volcanic edifice | Japan: accreted platforms of basalt, limestone and chert (Taira et al. 1983); subordinate calc-alkalic volcanic terranes (e.g., Nemura belt), and exotic seamounts; Middle to Late Cretaceous sub-aerial arc volcanic and plutonic rocks superimposed on accreted terranes (Taira and Tashiro 1987; Koido 1991). Southwest Pacific: early submarine basaltic to andesitic substrate overlain by basaltic to rhyodacitic subaerial volcanic rocks (Hamilton 1988; Macfarlane et al. 1988) |
| | **Sediment:** rapidly deposited conglomerate, wacke and siltstone spatially associated with and contiguous with calc-alkalic volcanic centres; oxide ironstone at top of volcanic sequences; subprovinces bordered by elongate sedimentary basins filled with turbiditic wacke, grainflow and conglomerate, minor ultramafic protrusions | Fore-arc prisms of greywacke turbidite (e.g., Shimanto Belt, Japan); sedimentary mélange with exotic fragments of volcanic rocks, and serpentinite (e.g., Maizuru belt, Watanabe et al. 1987) |
| Environments of volcanism and sedimentation | Early substrate of basaltic volcanic rocks: submarine and quiet, superimposed by increasingly emergent and explosive volcanic rocks, with final stage subaerial. Later volcanic rocks surrounded by ash flows, tuffs, proximal to distal sediment aprons; assemblages maybe separated by small-scale accretionary prisms | Volcanic assemblages predominantly tholeiitic to dacitic; calc-alkalic to alkalic volcanism only after coalescence of exotic fragments (Saito 1985; Zhang 1985) |
| Plutonism | Syn-volcanic, layered tholeiitic sills; subvolcanic calc-alkalic sills, plutons; syn-volcanic sheeted batholiths folded and reworked; post-volcanic peraluminous granitic batholiths and alkalic plutons | Tertiary calc-alkalic and peraluminous granitic plutons post-date accretion of exotic fragments (Kuroda and Yamada 1985; Saito 1985; Zhang 1985; Wallace et al. 1989; Osanai et al. 1991) |
| Structural evolution | Juxtaposition of originally separate volcanic assemblages and associated volcanically derived sediment wedges by thrusting and strike-slip shear zones; associated folding of sequences; disruption of collage by continued compression across arc and strike-slip motion | Thrust stacking of allochthonous assemblages (e.g., Hidaka belt, Horokanai ophiolite, Japan: Isozaki et al. 1990; Taira 1985; Taira et al. 1988); post-accretionary magmatism (Hibbard and Karig 1990); post-accretion strike-slip zones and faults (e.g., Median Tectonic Line: Saito 1985; Taira 1985; Taira et al. 1988) |
| Alteration and metamorphism | Hydrous alteration, subvolcanic and along assemblage contacts; early metamorphism except around syn-volcanic and pre-deformational plutons | Tilted sequences of arc exhibit granulite facies to greenschist facies; subduction complexes exhibit high pressure metamorphism (Taira and Tashiro 1987; Zhang 1985; Osanai et al. 1991) |
| Age range | Individual greenstone belts contain 300 million years of episodic volcanic activity | 300 million years of progressive accumulation and juxtaposition of arc-volcanic and sedimentary rocks (e.g., Japan: Taira et al. 1988; Zhang 1985) |
| Mineralization | Syn-volcanic hydrous alteration and VMS, replacement and vein systems; deep levels of gold and molybdenum; subvolcanic porphyry systems; ironstone deposition overlying felsic volcanic sequences, chromite in layered syn- to post-volcanic mafic to ultramafic intrusions; post-accretion gold-shear zone association; rare element pegmatites in S-type granitoids | Miocene, submarine exhalative (Kuroko) type syn-felsic volcanic base metal sulphide deposits (Ishihara 1978); Besshi type Cu-Zn deposits; bedded Fe-Mn deposits associated with basaltic sequences (Hutchinson 1983); Sn-W bearing ilmenite-granitoids (Ishihara 1978) |

## ORIGIN OF PLUTON-RICH AREAS AND SUBPROVINCES

The pluton-rich areas and discrete subprovinces, characterized by dominant sodic and potassic plutonism (Ayres 1978; Stone 1990; Thurston 1990a; *see* Thurston et al., this volume; Wilson 1971) may represent incomplete, oblique crustal sections through magmatic arcs (cf., Beard and Barker 1988; Thurston 1990b) separating subordinate supracrustal remnants (cf., Saleeby 1990). In any case, the generation of large volumes of primitive or near-primitive granitoid magma indicates that either large volumes of basaltic material or lower crustal rock were being melted (Beakhouse et al. 1989). Presumably, the melting and fractionation of such large rock volumes require an efficient mechanism of heat and fluid transfer, and crustal thickening such as that envisaged within and below modern volcanic arcs.

## SUBPROVINCE AGGREGATION

In previous sections of this chapter we tried where possible to describe the geological evolution of the Superior Province using nongenetic terminology. Before dealing with subprovince-scale analysis in this and following sections, the main findings of our greenstone belt-scale analysis are summarized, as follows:

1. Greenstone belts are tectonic collages of several depositional settings, including arc, oceanic, and unconformable, commonly fault-bounded environments.
2. Assemblages are elongate along each granite-greenstone subprovince and generally appear to young southward.
3. Greenstone belts and adjacent sedimentary subprovinces cannot be the result of simple rifting processes.
4. All granite-greenstone subprovinces display similar supracrustal rock types and depositional settings, physical and chemical evolution, structural patterns, scale, plutonism and geochronological diversity.
5. Variations within and between granite-greenstone subprovinces include geographic and secular variation in the distribution of komatiites, and geographic variability in the distribution of metamorphic grade and late-stage, unconformable, fault-bounded basins.
6. Although magmatic affinities, such as tholeiitic and calc-alkalic, in volcanism of both old and much younger belts or assemblages may be identical, alkalic volcanism tends to be more common in later assemblages. As well, there is an apparent restriction of continental platformal sedimentary rocks to old assemblages. Both of these observations indicate secular change.

We submit that the striking similarities between the stages of geological development of the Superior Province and the geology of modern island arc complexes (Isozaki et al. 1990; Osozawa 1988), and the useful application of terrane analysis in all of the granite-greenstone subprovinces (*see* Thurston et al., this volume), justifies the usage of plate tectonic analogues (Barley et al. 1989; Barrie and Davis 1990; Williams 1990). After briefly reviewing such parallels (*see* Tables 25.2 and 25.3), we will discuss the evolution, role and timing of the accretionary prisms, the processes and timing of subprovince aggregation, and compare the geophysical signatures of subprovinces and their boundaries with those of terranes in modern orogens.

### Justification of the Arc-Arc Accretion Model

The similarities between Archean supracrustal assemblage types and the lithotectonic units in modern continental and island arcs and related tectonic environments are listed in Table 25.2. In Table 25.3, Archean greenstone belts, pluton-rich and sedimentary subprovinces are compared with modern tectonic environments in terms of scale, time, rock type, magmatic, geochemical and structural evolution, and mineralization. This comparison strongly suggests that modern arc-arc accretion phenomena (e.g., Isozaki et al. 1990) yield a geological record little different from that seen in the Superior Province.

Rift models may explain late-stage clastic and volcanic deposits associated with some greenstone belts (e.g., Timiskaming assemblage, Abitibi Subprovince, *see* Jackson and Fyon, this volume; Whim Creek belt, Pilbara Block, Australia, Barley 1987). However, extensional zones such as continental and midoceanic rifts or wrench faults, cannot be used to explain the geology of the bulk of greenstone belt evolution (Williams 1990) because:

1. In many instances, there is little or no field, geochemical or geophysical evidence of basement rocks.
2. The geology is more complex, episodic and long-lived than can reasonably be explained by processes associated with rift, wrench-fault or plume-generated zones.
3. The relative timing of the complex volcanic, sedimentary and plutonic rocks in greenstone belts cannot be explained by simple intracontinental fault-related models.
4. The inferred sources and geochemical signatures of magmatic rocks, and the efficiency of and degree to which the inferred source rocks have been melted, are not characteristic of rifts or wrench-faulted zones.
5. Evidence is accumulating (*see* Stott and Corfu, this volume) that greenstone belts and subprovinces are asymmetric features with respect to age and structure, when compared to similar belts in Phanerozoic orogens.

Large-scale elongate belts of contrasting rock associations constitute the major evidence of accretionary processes, supported by structural data of the Superior Province. Together with the distribution of folds, faults and patterns of strain, data from all subprovinces in Ontario seem consistent with large-scale crustal shortening. However, both active island arcs and postcollisional orogens are commonly also sites of extension (Charlton 1991; Hamilton 1988; Honza 1991). Large areas of granitoid rocks within the Superior Province display flat-lying and strongly developed mineral foliations, extensional lineations and extensional shears, which suggests that granitoid terrains may have undergone synorogenic extension (e.g., Wawa gneiss terrane , Moser et al. 1991). This extension may be due simply to relaxation on uplift, and erosion resulting from tectonic thickening; alternatively, the work of Moser et al. (1991) shows that this extension was synchronous with, and therefore perhaps complementary to, features typical of shortening in adjacent supracrustal-dominated regions.

## Subprovince Boundaries

Subprovince boundaries are typically structural and metamorphic transition zones between subprovinces with contrasting lithological character. Variable amounts of late-stage strike-slip faulting and emplacement of intrusions occurred along these boundaries, for example, granitoid intrusions along the western section of the Uchi–English River subprovince boundary (*see* Breaks, this volume), in part obliterating earlier structural features including those indicative of dip-slip faulting and recumbent folding.

Where subprovince boundaries have not been affected by later strike-slip motions, thrust faults with a north over south shear-sense and recumbent folding have been inferred (e.g., Wawa–Quetico and Wabigoon–Quetico subprovince boundaries, *see* Williams, this volume). These "thin-skinned" dip-slip motions and accompanying upright folding were followed by "thick-skinned" thrust faulting, perhaps interspersed with greenstone belt-parallel strike-slip motions, such as in the Beardmore–Geraldton Belt at the Wabigoon–Quetico (*see* Williams, this volume) and Uchi–English River subprovince boundaries (Stone 1981; Kamineni et al. 1990; *see* Stott and Corfu, this volume). These motions have been recognized as transpressional (Sanborn-Barrie et al. 1990; Stott et al. 1987), a process well known from modern orogens.

Not all subprovince boundaries show the same north over south shear-sense. For example, an opposite shear-sense has been documented along the eastern part of the Uchi–English River subprovince boundary (*see* Breaks, this volume). We have no explanation for this feature, but note that such tectonic complexity is typical of modern island and continental arcs where evolving triple junctions (Teng 1990), change in direction of subduction or postcollisional extension, such as that associated with the Banda–Australian collisional zone (Charlton 1991), create regional changes in shear-sense .

Subprovince boundaries commonly seem to change character along their length because of postaccretion faulting. In the southern Superior Province, not all subprovince boundaries are original; some have been modified or cut by strike-slip faults such as the Wabigoon fault (*see* Blackburn et al., this volume) and the Quetico Fault (*see* Williams, this volume). The Wabigoon–Quetico subprovince boundary is a lithological and structural transition zone in the Beardmore–Geraldton Belt (Devaney and Williams 1989) but farther west, in the Atikokan area, the Quetico Fault juxtaposes contrasting lithological and structural characteristics across a well-developed zone of strike-slip faulting.

## Accretionary Prism Evolution

Calc-alkalic volcanism in granite-greenstone subprovinces ended at about 2.70 Ga, at about the time of their amalgamation with intervening volcanic accretionary prisms and pluton-dominated subprovinces. The formation and evolution of large-scale accretionary prisms marked the onset of a major change in the scale of the dominant tectonic processes in the Superior Province (Williams 1990). Prior to major accretionary prism development, tectonism in granite-greenstone subprovinces involved an assemblage-scale interaction measurable in tens rather than hundreds of kilometres; however, the largest intragreenstone sedimentary assemblages, such as the Gargantua assemblage in the Wawa Subprovince (*see* Figure 25.4; *see also* Williams, Stott, Heather et al., this volume) and the Billett assemblage in the Uchi Subprovince (*see* Stott and Corfu, this volume), may foreshadow the later development of large-scale sedimentary prisms. The onset of subprovince coalescence produced clastic sediment-dominated tectonic units up to a hundred kilometres wide and over a thousand kilometres long, on a scale comparable to those typical of modern continental and island arc systems (Hamilton 1988).

The sediments in accretionary prisms are dominantly derived from a volcanic, rather than a plutonic source, as indicated by ubiquitous clasts of plagioclase and plagioclase-phyric rock within low-grade metasedimentary rocks (Ayres 1983; *see* Breaks, this volume; van de Kamp and Beakhouse 1979; *see* Williams, this volume). The occurrence of 3 billion-year-old zircon grains in the metasedimentary rocks of the Quetico Subprovince indicates a possible provenance in the ancient plutonic rocks of the Atikokan region (Davis et al. 1990). Yet, igneous zircon grains of 2.70 Ga age are common within Quetico Subprovince metasedimentary rocks (Davis et al. 1990). The adjacent granite-greenstone Wabigoon Subprovince, considered to be the source area for the Quetico, contains little evidence of volcanic rocks of this age. Indeed, the youngest late-stage volcanism preserved on the south flank of the granite-greenstone subprovince has been dated at 2720 to 2710 Ma (*see* Blackburn et al., this volume). We infer that calc-alkalic volcanism of 2700 Ma age would likely have continued up until the time of subprovince collision at about 2690 Ma. Perhaps late-tectonic, subaerial, calc-alkalic (ca. 2700 Ma) and shoshonitic (ca. 2689 Ma) volcanism on the southern flanks of the granite-greenstone subprovinces was mostly removed tectonically, or by erosion (cf., Ayres 1983).

## Evidence of Ensialic Basins

A small number of synorogenic to late orogenic basins, the so-called "late unconformable basins", occur within the Superior Province. Some, but not all, are associated with strike-slip faults; some may display shoshonitic magmatism; all seem to have developed between the major stage of recumbent fabric formation but predate the transpressive and shear-related deformation. A few examples are cited:

1. Shoshonitic rocks, such as the Bijou Point complex (~2.70 Ga), are spatially associated with strike-slip faults (the Bear Head fault and North Spirit shear zones; *see* Thurston et al., this volume). Felsic volcanic rocks of the North Spirit assemblage, 2735 Ma in age, unconformably overlie much older crust; nearby, at Bijou Point, a contemporaneous shoshonitic intrusion of a type similar to that found within the Timiskaming assemblage (*see* Jackson and Fyon, this volume), has been dated at 2731 Ma (Corfu and Wood 1986). The North Spirit assemblage is characterized by a smaller proportion of mafic metavolcanic rocks, and more abundant felsic metavolcanic rocks and related epiclastic units, relative to other older volcanic assemblages. The units contain an inherited zircon component dated at ~2862 Ma (Corfu and Wood 1986), implying eruption through or partial derivation from older sialic crust.
2. The rocks of the Timiskaming assemblage (~[illegible]) postdate subprovince aggregation, because the [illegible] they contain overlap subprovince boundaries. The Timiskaming and Shebandowan assemblages have hitherto been interpreted as resulting from strike-slip (Cooke and Moorhouse 1969; Hyde 1980) or pull-apart (Thurston and Chivers 1990) processes (cf., Christie-Blick and Biddle 1985).
3. Sedimentary rocks without coeval volcanic rocks or derived coarse volcanic detritus, such as the Three Nations, Garrison and Hearst assemblages in the Abitibi Subprovince also occur (*see* Jackson and Fyon, this volume).

Criteria to distinguish "Timiskaming-type" clastic rocks (*see* Figure 25.5) from older, coarse-grained synvolcanic sedimentary assemblages are not well established, but the following have been locally used to define "Timiskaming-type" rocks:

1. commonly coarse-grained, of local provenance
2. at least in part unconformable on older greenstone belt rocks
3. the sedimentary rocks are commonly, but not universally, intercalated with alkalic volcanic rocks and derived clasts
4. some, but not all "Timiskaming-type" sedimentary rocks are spatially associated with major strike-slip or complex fault zones
5. ensialic supracrustal rocks escaped an early deformation that affected subjacent greenstone belt rocks

However, it should also be noted that:

1. not all sedimentary rocks in ensialic basins are coarse-grained
2. not all contain associated alkalic volcanic rocks
3. not all are clearly associated with late-tectonic fault zones
4. most have not had the benefit of structural analysis, so that time relationships to deformation are largely unknown

Some ensialic rock sequences exhibit 1 or 2 aspects of the critical field evidence suggestive of an origin similar to that proposed for the Timiskaming assemblage, but also exhibit field relationships suggestive of a different origin. We conclude that the spectrum of sedimentary rocks that developed near the end of granite-greenstone subprovince development largely defies uniform distinction into discrete members, and that tectonic processes operating during the final stages of granite-greenstone subprovince development and subprovince collision were regionally diachronous.

## Alkalic Magmatism

Neoarchean alkalic intrusive rocks are rare; examples occur at: Poohbah Lake (Quetico Subprovince, *see* Williams, this volume), Sturgeon Lake (Wabigoon Subprovince, *see* Blackburn et al., this volume; *see also* Sage, this volume), Dickenson Lake (Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume), Springpole Lake (Uchi Subprovince, *see* Stott and Corfu, this volume; *see also* Barron et al. 1989) and Wapikopa Lake (Sachigo Subprovince, *see* Sage, this volume).

Field relationships and geochronological data indicate that the saturated to undersaturated syenitic pluton at Poohbah Lake intruded ductile metasedimentary rocks during high-grade metamorphism and deformation. This intrusion is not spatially associated with regional-scale faults, but occurs within rocks newly accreted as thrust fault-bounded units and may be the result of vestigial subduction-related magmatism of an extremely fractionated type, related to the sanukitoid suite (Stern et al. 1989).

Foliated to undeformed plutons of syenitic composition on Sturgeon Lake are an example of Neoarchean alkalic bodies that are syn- to posttectonic. Similar relationships hold for the Springpole and Wapikopa occurrences. Both the Sturgeon Lake and Springpole occurrences have been dated at prior to the arc-arc accretion event, so they may be associated with an earlier brittle regime postdating the aggregation and early deformation of assemblages.

These bodies may be penecontemporaneous with Neoarchean calc-alkalic volcanism, and some may relate to the reactivation of nearby fault boundaries between tectonic assemblages and to vestigial magmatism associated with progressive crustal stabilization.

## Use of a Cladogram in Tectonic Analysis

The cladogram (Young 1986), shown in Figure 25.7, summarizes the complex space-time relationships amongst tectonic assemblage, terrane and superterrane subdivisions proposed to constitute the Superior Province in Ontario. A cladogram is a branching diagram, commonly used in biosystematics to reconstruct the phylogeny of plant or animal species; it was first used in tectonic analysis by Young (1986) and subsequently by Hoffman (1989) to summarize the accretionary history of the North American craton. Cladistics is a useful method to analyze the sequence of events, the interrelationships between crustal blocks, and the timing of assembly of these crustal pieces to form an Archean craton. A cladogram can be organized to resemble a hierarchical time-space chart, with branching lines converging from tectonic assemblages or terranes toward a common time line that begins at the time of formation of a mass of older crust. The converging branches represent accretionary events or other crustal additions.

## Construction of the Superior Province

The Archean Superior Province is viewed as having been constructed in 2 stages: 1) the internal construction of subprovince-scale superterranes or composite terranes, and 2) the assembly of these superterranes to form the Superior Province during the Kenoran Orogeny at about 2.7 Ga. The assembly of different tectonic components of an Archean subprovince is constrained by different magmatic and deformational events that can be dated.

The cladogram in Figure 25.7 illustrates 1 possibility: that the oldest rocks in northwestern Ontario were initially part of a single terrane (proto-North Caribou terrane) that split up to form smaller terranes. Crust was added to each of these smaller terranes during later episodes of growth, by accretion or by additions along microcontinental margins. These composite terranes converged and amalgamated to form a single craton. In northwestern Ontario, the oldest supracrustal rocks, which occur in the North Caribou terrane (Thurston and Davis 1991) of the Sachigo Subprovince, formed the core of a microcontinent to which younger Archean crust was added at episodic intervals. The crustal additions are terranes composed of tectonic assemblages, some of which are represented on the cladogram. Several terranes of contrasting ages have been interpreted by Thurston and Davis (1991) to form the Sachigo Subprovince of Card and Ciesielski (1986). The Uchi Subprovince, along the southern flank of the Sachigo Subprovince, based upon revised nomenclature suggested in this chapter, was formed during separate episodes when crust was added onto the southern flank of the North Caribou terrane (*see* Stott and Corfu, this volume) and was deformed and intruded by numerous plutons during an inferred collision with other terranes from the south. The Wabigoon and Wawa subprovinces (e.g., Williams, Stott, Heather et al., this volume) are composite terranes, and evidence presented by Corfu and Stott (1986), Stott et al. (1987) and Davis et al. (1989) indicates that the amalgamation of composite terranes progressed southward (using present day directions). This event, known as the Kenoran Orogeny, was a rapid, progressive deformation and plutonism during terrane amalgamation, and represents the second stage of Superior Province construction.

## Summary of Events Leading up to Arc-Arc Accretion and Coalescence

The stages which culminated in the coalescence of subprovinces into the Superior Province are summarized as follows:

1. prior to 3.0 Ga: fragmentation of sialic crust seems likely, since many fragments are now preserved as old relics in many granite-greenstone subprovinces (*see* Figure 25.7)
2. 3.10 to 2.70 Ga: progressive aggregation of greenstone assemblages into linear belts upon which were superimposed calc-alkalic volcanic centres and major batholithic complexes. Several stages of aggregation and calc-alkalic magmatism are evident in the Sachigo (2.871 Ga) and Wawa (2.89, 2.75 and 2.70 Ga) subprovinces, as is a southward progression of assemblage accretion in each subprovince. A speculative series of accretionary events has been catalogued for the Sachigo Subprovince (*see* Thurston et al., this volume).
3. 2.70 to 2.69 Ga: consequent erosion of prior and newly formed volcanic material sheds detritus into neighbouring sedimentary basins
4. 2.69 Ga: progressive deformation and tectonic thickening of sedimentary rock, magmatic thickening of granite-greenstone subprovinces, underthrusting of granite-greenstone subprovinces beneath the leading edges of sedimentary prisms (e.g., proto-Wawa beneath proto-Quetico); progressive southward decrease in age of accretion of each subprovince to its southerly neighbour
5. 2.69 to 2.65 Ga: local thickening of sedimentary welt and adjacent greenstone subprovinces, melting and high-grade metamorphism induced where thickening is substantial; local extensional features, including the formation of fault-bounded basins and subsequent incorporation of this supracrustal rock into middle crustal levels
6. 2.65 to 2.45 Ga: continued convergence of subprovinces, subprovince-parallel slip and continued shortening by a combination of underthrusting and lateral escape; gradual cooling of midcrustal levels

# USE OF GEOPHYSICAL DATA IN TECTONIC ANALYSIS OF THE SUPERIOR PROVINCE

## Geophysical Signature of Greenstone Belts and Subprovinces

Aeromagnetic, gravity, seismic and radiometric studies in the last 3 decades have complemented the geological sub-division of the Superior Province into subprovinces; each subprovince displays a distinct mix of geophysical properties (Hall and Brisbin 1982). The data have also helped to unravel the internal structure of greenstone belts (Figure 25.8a) and subprovinces (Figure 25.8b). The salient points made in the detailed presentation and analysis of geophysical data for the Uchi Subprovince (*see* Stott and Corfu, this volume) have significance for the Superior Province as a whole and are reiterated here.

## Aeromagnetic Data

Aeromagnetic distinction between the elliptical to elongate patterns of granite-greenstone subprovinces and the more regular and less contrasting anomalies characteristic of sedimentary rock-dominated subprovinces, for example, the English River Subprovince (Urquhart and West 1979), is relatively straightforward (Figure 25.9). Within granite-greenstone subprovinces in northern regions not dominated by later diabase dike swarms, units of iron formation form a dominant constituent of the magnetism of supracrustal rocks and are distinct from the inhomogeneous mafic volcanic rocks with a low magnetic susceptibility; these in turn are distinguishable from the higher magnetic susceptibilities of bodies of tonalite and granodiorite to granite (McGrath and Hall 1969), some of which are prominently magnetic (e.g., Williams suite, Uchi Subprovince, *see* Stott and Corfu, this volume). In southern regions, especially in the Abitibi Subprovince, the Matachewan diabase dike swarm largely precludes geophysical subdivision of granitic terrain.

Magnetic anomaly patterns in greenstone belts have helped to distinguish assemblages from each other. Discordant magnetic trends within belts commonly meet along linear zones where 1 set of trends is transected by those of a neighbouring assemblage (e.g., Pickle Lake greenstone belt, Uchi Subprovince, *see* Stott and Corfu, this volume). These zones are assemblage boundaries and, where verified in the field, correspond with lithological, age or structural contrasts.

At and near some margins of both sedimentary and granite-greenstone subprovinces, magnetic response is intense, strongly linear, and correlates with the differences in structural style and metamorphic mineralogy enumerated earlier in this discussion (e.g., Wawa–Quetico subprovince boundary, Williams and Breaks 1990).

In some cases, the sense of displacement on large-scale shear zones and faults (e.g., Bear Head fault zone, *see* Thurston et al., this volume; *see also* Stott and Corfu, this volume) can be determined from magnetic signatures (Osmani and Stott 1988).

## Gravity Data

Newly compiled Bouguer anomaly maps of Ontario (*see* Maps 2592 to 2595, map case) show that positive gravity anomalies are generally associated with greenstone belts (e.g., Michipicoten greenstone belt) and dense, highly metamorphosed rocks (e.g., granulite facies rocks in the English River Subprovince and Kapuskasing Structural Zone). Distinction between volcanic and sedimentary rocks in greenstone belts is possible only where detailed data collection allows the production of second derivative maps of the Bouguer gravity data (e.g., Gupta and Barlow 1984; Gupta et al. 1982; Gupta and Wadge 1986; Szewczyk and West 1977).

Gravity data for several greenstone belts (*see* Figure 25.8a) indicate that the belts are rarely more than 5 km in vertical extent, and commonly only 2 to 3 km (Gupta et al. 1982; Gupta and Wadge 1986; Dusanowskyj and West 1976). Similarly, Gupta and Osmani (Ontario Geological Survey, personal communication, 1990) have shown that the vertical extent of the Abitibi greenstone belt varies from 3 to 12 km. Gravity data permit a "thin-skinned" model of thrust-stacking and imbrication of contrasting assemblages.

In a study of paired gravity anomalies observed at the Uchi–English River subprovince boundary (*see* Figure 25.8b), Thomas (1989) proposed that the subduction polarity there was south dipping. This inference conflicts with most geological data, such as the southward-younging assemblages within the Uchi Subprovince and occurrence of an east-trending train of late-tectonic batholiths in the latter (e.g., Williams suite, *see* Stott and Corfu, this volume).

## Seismic Data

The acquisition of seismic refraction and reflection data has produced a wealth of information on vertical and horizontal velocity distributions within the earth's crust, and the depth, orientation and thickness of reflecting geological units.

Seismic refraction data have allowed the determination of crustal thickness and structure along a transect stretching from Kenora to Red Lake (*see* Figure 25.8b), crossing several subprovinces of the Superior Province (Hall and Brisbin 1982). This data also indicates that subprovince boundary zones are coincident with abrupt changes in crustal thickness (e.g., Uchi–English River and English River–Winnipeg River subprovince boundaries, Figure 25.8b).

Local seismic refraction data within greenstone belts indicate that some granitoid plutons are relatively thin subhorizontal sheets, rather than vertical-sided batholiths extending to great depth; this is consistent with gravity data suggesting that greenstone belts do not have thick roots, but are masses only about 5 km thick (e.g., Dusanowskyj 1976; Gupta et al. 1982; *see* Figure 25.8a).

Seismic reflection profiles have been acquired for 2 major structural regions in the Superior Province: the Kapuskasing Structural Zone (Percival et al. 1989), and the Abitibi Subprovince (Jackson et al. 1990). Although there are difficulties in matching surface geology with near-surface reflectors, one of the most intriguing aspects of the 2 studies is the identification of subhorizontal to shallowly dipping seismic reflectors at middle to upper crustal levels. These are still of unknown origin; they might be lithological units, structures, or an interbanding of contrasting rock types. Seismic reflection studies at this regional scale indicate that the crust is layered, but the nature of the seismic reflectors is as yet undetermined. Significantly, reflectors of similar scale and geometry have been noted in refraction

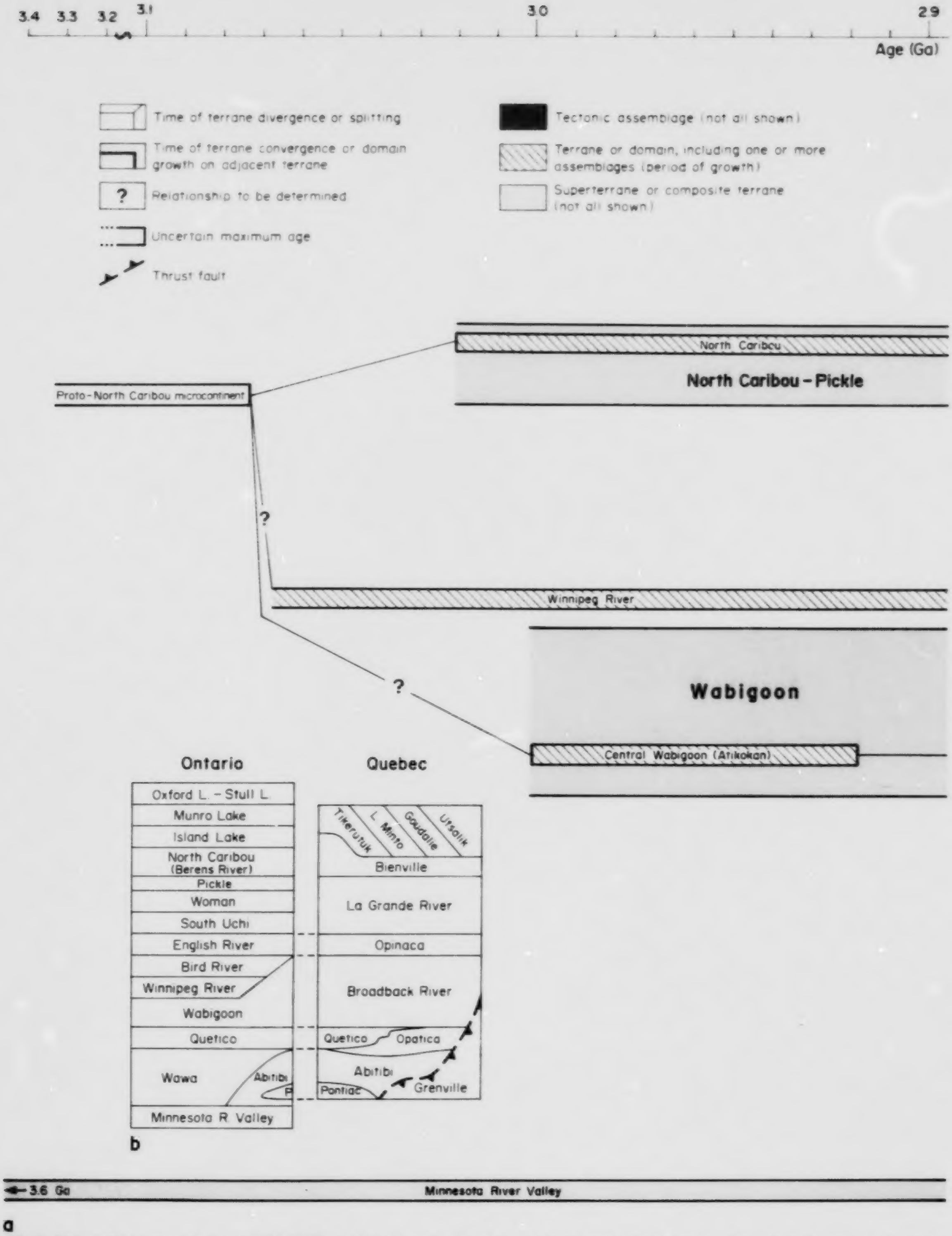

3.4
3.3
3.2
3.1
3.0
2.9
Age (Ga)
Time of terrane divergence or splitting
Time of terrane convergence or domain growth on adjacent terrane
?
Relationship to be determined
Uncertain maximum age
Thrust fault
Tectonic assemblage (not all shown)
Terrane or domain, including one or more assemblages (period of growth)
Superterrane or composite terrane (not all shown)
North Caribou
North Caribou - Pickle
Proto-North Caribou microcontinent
?
Winnipeg River
?
Wabigoon
Central Wabigoon (Atikokan)
Ontario
Quebec
Oxford L. - Stull L.
Munro Lake
Island Lake
North Caribou (Berens River)
Pickle
Woman
South Uchi
English River
Bird River
Winnipeg River
Wabigoon
Quetico
Wawa
Abitibi
P
Minnesota R. Valley
Tikerutuk
L. Minto
Goudalie
Utsalik
Bienville
La Grande River
Opinaca
Broadback River
Quetico
Opatica
Abitibi
Pontiac
Grenville
b
3.6 Ga
Minnesota River Valley
a

**Figure 25.7. a)** Cladogram (cf., Young 1986; Hoffman 1989) illustrating known and inferred spatial and temporal relationships of tectonic units within the Superior Province, emphasizing that part within Ontario, and the sequence of major accretionary events leading to the coalescence of subprovinces. Some of the age constraints are based on U-Pb zircon age determinations provided by F. Corfu and D. Davis of the Royal Ontario Museum. **b)** Schematic diagram indicating the spatial relationships of the major accreted terranes and a proposed correlation of units between the northwestern and northeastern parts of the Superior Province.

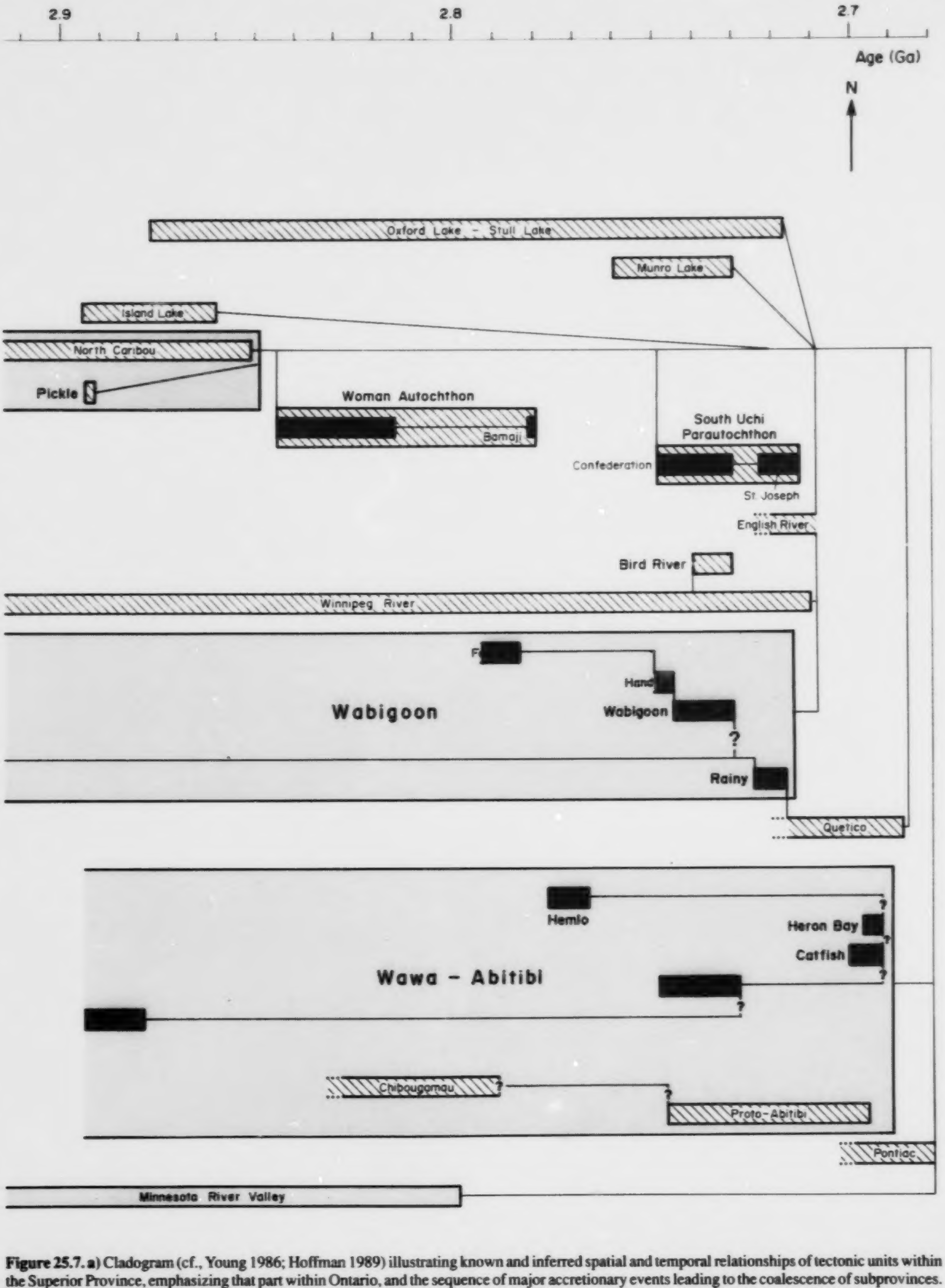

2.9
2.8
2.7
Age (Ga)
N
Oxford Lake – Stull Lake
Munro Lake
Island Lake
North Caribou
Pickle
Woman Autochthon
Bamaji
South Uchi
Parautochthon
Confederation
St. Joseph
English River
Bird River
Winnipeg River
F
Hand
Wabigoon
Wabigoon
?
Rainy
Quetico
Hemlo
Heron Bay
Catfish
Wawa – Abitibi
?
Chibougamau
Proto-Abitibi
Pontiac
Minnesota River Valley

**Figure 25.7. a)** Cladogram (cf., Young 1986; Hoffman 1989) illustrating known and inferred spatial and temporal relationships of tectonic units within the Superior Province, emphasizing that part within Ontario, and the sequence of major accretionary events leading to the coalescence of subprovinces. Some of the age constraints are based on U-Pb zircon age determinations provided by F. Corfu and D. Davis of the Royal Ontario Museum. **b)** Schematic diagram indicating the spatial relationships of the major accreted terranes and a proposed correlation of units between the northwestern and northeastern parts of the Superior Province.

data recorded in an actively accreting environment in southern Alaska (Fuis et al. 1991).

A relatively new seismic technique records the modification of natural earthquake energy, revealing splitting and velocity contrasts of shearwaves, brought about by structural anisotropy in the lower lithosphere and upper mantle beneath sections of the Superior Province (Silver and Chan 1988). Continental-scale anisotropy is known from modern orogens to be an artifact of lithospheric strain during orogenesis; what is novel about this study is the evidence of a corresponding anisotropy in the subcontinental mantle beneath part of the Superior Province. The trend of the mantle anisotropy beneath the Uchi Subprovince (for a fuller discussion of technique and results, *see* Stott and Corfu, this volume) is parallel to the subprovince boundary and the major assemblage boundaries in that region. It appears that the mantle anisotropy remained intact after 2.7 Ga.

Analysis of seismic energy from distant earthquakes using the tomographic technique allows the measurement of lithospheric thickness seismic velocity structure of the base of the lithosphere. The Superior Province is underlain by an especially thick lithosphere compared to the rest of North America (Grand 1987). The significance of this observation plus the preservation of 2.7 billion-year-old mantle anisotropy is that the large Archean lithospheric thickness may in part correlate with crustal stability and is not the product of later crustal thickening.

## Radiometric Data

Within the Superior Province, gamma ray data and its conversion to uranium, thorium and equivalents (e.g., Geological Survey of Canada 1966) reveal detail in the structure and composition of the granitoid regions that make up much of the granite-greenstone subprovinces. In particular, they allow distinction between the relatively nonradioactive tonalite-granodiorite suite, and younger, more radioactive granitic rocks, which when used in conjunction with aeromagnetic data allowed a refinement of the boundaries of the granitoid regions in the northwestern Superior Province (*see* Thurston et al., this volume).

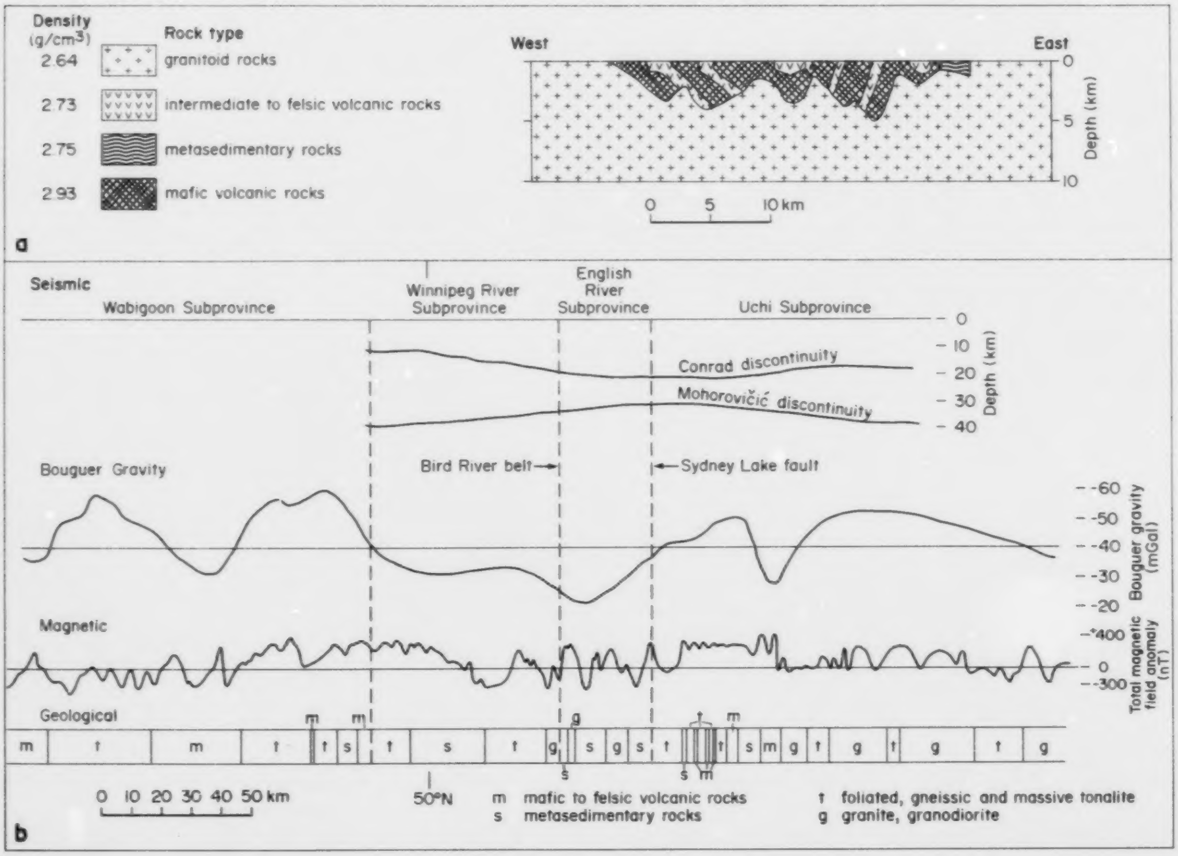

Density (g/cm3)
Rock type
2.64 granitoid rocks
2.73 intermediate to felsic volcanic rocks
2.75 metasedimentary rocks
2.93 mafic volcanic rocks
West
East
0
5
10
Depth (km)
0 5 10 km
a
Seismic
Wabigoon Subprovince
Winnipeg River Subprovince
English River Subprovince
Uchi Subprovince
0
10
20
30
40
Depth (km)
Conrad discontinuity
Mohorovičić discontinuity
Bouguer Gravity
Bird River belt
Sydney Lake fault
-60
-50
-40
-30
-20
Bouguer gravity (mGal)
Magnetic
400
0
-300
Total magnetic field anomaly (nT)
Geological
m t m t t s t s t g s g s t s m g t g t g t g
0 10 20 30 40 50 km
50°N
m mafic to felsic volcanic rocks
s metasedimentary rocks
t foliated, gneissic and massive tonalite
g granite, granodiorite
b

**Figure 25.8.** Geophysically modelled cross sections: **a)** through a typical greenstone belt and its associated granitoids, as determined from detailed gravity observations (*modified from* Figure 13, Gupta et al. 1982); and **b)** seismic, gravity, magnetic and geological sections along long. 94°W, from the northern edge of the Wabigoon Subprovince, through the Winnipeg River and English River subprovinces, into the southern part of the Uchi Subprovince (*modified from* Hall and Hajnal 1969, and Maps 2542, 2585, 2593, map case).

## Paleomagnetic Data

Stott and Corfu (this volume) fully discuss paleomagnetic data for the Uchi Subprovince. The significant points are that recent work (Hale and Lloyd 1990) demonstrate that the Archean apparent polar wander path needs revision and that intracratonic movement within the Superior Province after 2.7 Ga was minimal. In addition, Hale and Lloyd (1990) inferred from detailed paleomagnetic data that individual plutons in blocks separated by major faults have rotated clockwise prior to 2.7 Ga.

## Correlation of Geophysical and Geological Data

Magnetic data help to define assemblage boundaries in greenstone belts. By tracing individual magnetic units within assemblages in greenstone belts, it has been shown that greenstone belts are made up of disparate, commonly discordant rock masses with contrasting magnetic responses (*see* Jackson and Fyon, this volume). The field recognition of assemblages as shear-bounded panels may imply the presence of a "master" décollement. Geophysical evidence for such a relatively shallow, flat-lying structure comes from both gravity and seismic reflection profiling (Jackson et al. 1990).

# TECTONIC HISTORY AFTER SUBPROVINCE COALESCENCE

Geological evolution of the Superior Province did not end with completion of the convergence and amalgamation of the island arc-like subprovinces. A protracted period of faulting and dike emplacement continued throughout the

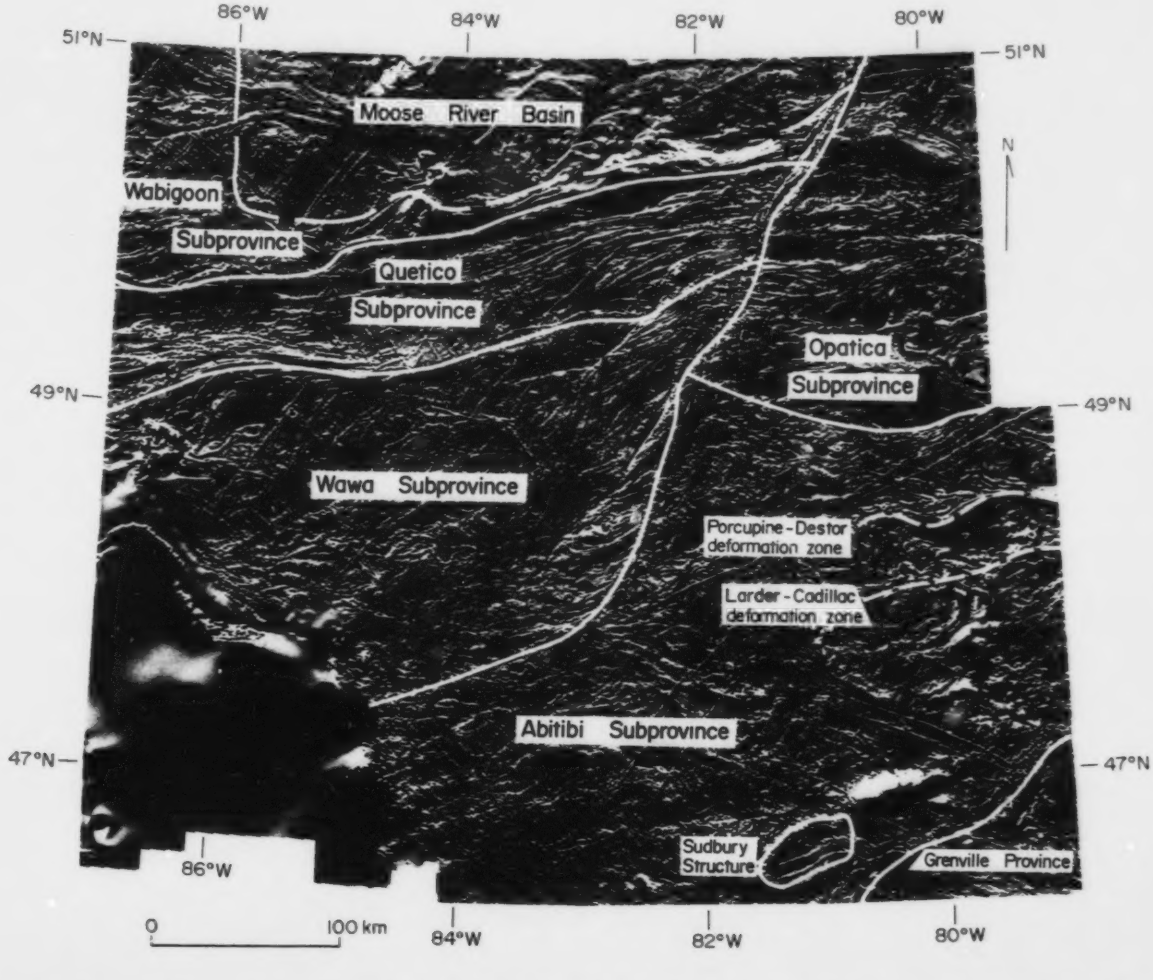

86°W
84°W
82°W
80°W
51°N
51°N
Moose River Basin
N
Wabigoon
Subprovince
Quetico
Subprovince
Opatica
Subprovince
49°N
49°N
Wawa Subprovince
Porcupine - Destor deformation zone
Larder - Cadillac deformation zone
Abitibi Subprovince
47°N
47°N
Sudbury Structure
Grenville Province
86°W
0
100 km
84°W
82°W
80°W

**Figure 25.9.** Shaded image of the total magnetic field intensity of the eastern part of the Superior Province. The map includes parts of the Abitibi, Quetico and Wabigoon subprovinces. The greenstone belts are characterized by short-range order and a variety of intensities, whereas the sedimentary subprovince shows medium-range order with only rare areas of high intensity arising from the presence of peraluminous plutons and areas of migmatization in the interior of the subprovince. Subprovince boundary shear zones are also quite evident.

Proterozoic, in part related to the formation of the Penokean Fold Belt to the south and ending with the development of the Midcontinent Rift during the Mesoproterozoic (*see* Sutcliffe, this volume).

## Late Stage Faulting and Dike Intrusion

Subprovince boundaries and ductile deformation zones contained within the subprovinces became progressively affected by brittle deformation along the major strike-slip faults that developed after subprovince coalescence (Osmani and Stott 1988; *see* Thurston et al., this volume) and which continued during the Proterozoic (Kamineni et al. 1990; Peterman and Day 1989). Isotopic dating of pseudotachylyte by Rb-Sr methods (1947±23 Ma, Peterman and Day 1989; 2183±74 Ma, Kamineni et al. 1990) indicates that Paleoproterozoic activity is recorded on faults. Systems of northeast-trending (sinistral) and east-southeast-trending (dextral) strike-slip faults record brittle responses to stresses similar to those that caused the previous coalescence.

Several separate swarms of diabase dikes, ranging in age from 2.45 to 1.14 Ga (*see* Osmani, this volume), cut all structural features other than the late-stage faults described above. Basaltic magmas were intruded along fissures, starting in Paleoproterozoic time, but the most volumetrically significant swarm formed along a direction which had formerly been parallel to the maximum principal compressive stress direction at the end of the Archean (*see* Osmani, this volume). This major swarm, the Matachewan, has been ascribed to early Paleoproterozoic magmatism associated with crustal extension orthogonal to the east-trending Huronian passive margin (*see* Bennett et al., this volume; Heaman 1988).

## Archean Control of Proterozoic Sedimentation and Magmatism

At about 2450 Ma, Proterozoic sedimentation and associated basaltic magmatism in what is now the Southern Province (*see* Bennett et al., this volume) began along what is now the southern border of the Superior Province. It is not yet clear how much of the Superior Province occurs in modified form beneath and to the south of this Paleoproterozoic basin. Bennett et al. (this volume) describe the control of the east-trending Murray Fault System on the distribution of preserved magmatism and sedimentation. The Murray Fault System is a major dextral strike-slip structure parallel to the Larder–Cadillac and Porcupine–Destor deformation zones, and may coincide with the southern boundary of the Abitibi Subprovince in Ontario. Paleoproterozoic magmatism, in the form of the north-trending Matachewan diabase dike swarm (*see* Osmani, this volume), occupies late-stage fractures and faults produced during the final stages of brittle behaviour soon after the transpressive deformation of the Superior Province. In part, it is these faults that control the distribution, facies and preservation of Paleoproterozoic sedimentary rocks in the Cobalt Embayment (*see* Bennett et al., this volume).

## The Kapuskasing Structural Zone

The Kapuskasing Structural Zone (KSZ) has been interpreted to be an intracratonic uplift along a major listric fault (Percival and Card 1983, 1985; Leclair 1990; Percival 1990). In this section, we have adopted a modified interpretation of this hypothesis; we emphasize that rocks exhibiting high metamorphic pressures and unusual structures are not unique to this zone, but occur in several locations, specifically at subprovince boundaries and on the hanging walls of other major sinistral oblique-slip faults (Wawa Subprovince, *see* Williams, Stott, Heather et al., this volume; Winnipeg River Subprovince, *see* Beakhouse, this volume; English River Subprovince, *see* Breaks, this volume).

The KSZ is a zone of uplift representing exposure of midcrustal rocks which have been subjected to metamorphic pressures of 0.5 to 1 GPa (Percival 1990; Percival and Card 1983); however, the following observations are also pertinent to this interpretation:

1. The regional distribution of rock types within the KSZ is a pattern of contrasting tracts correspondingly rich in mafic or sedimentary rocks (*see* Map 2543, map case); rocks exposed in the southern part of the uplift are similar to those found along a cross section through the boundary between the Wawa and Quetico subprovinces (Williams and Breaks 1990). Similarly, rocks in the northern part of the KSZ are comparable to those documented from the Wabigoon–Quetico subprovince boundary.
2. Tectonic fabrics, such as well-developed foliations and lineations within the KSZ, are continuous with, and parallel to those seen in other regions of high metamorphic grade, such as along the contact between the Wawa and Quetico subprovinces to the west (Williams and Breaks 1990).
3. The range of metamorphic pressures and temperatures determined from rocks within the KSZ are higher than those in adjacent rocks; however, temperatures of ~900°C and pressures on the order of 0.5 to 0.7 GPa have been determined from the boundary between the Wawa and Quetico subprovinces in the Manitouwadge region, which is unaffected by major faults (Y. Pan, University of Western Ontario, personal communication, 1991); metamorphic pressures of 0.5 to 0.6 GPa have been determined from granite-greenstone terrane in the northern part of the Wawa Subprovince (Pan 1990).

We note that subprovince boundaries, such as that documented from the Manitouwadge region, display regional progressions of rock type and structure similar to those described within the KSZ, though with less extreme metamorphic grades (H.R. Williams 1991; Williams and Breaks 1990). Perhaps the KSZ contains disrupted parts of the boundaries between the Wawa and Quetico and Wabigoon and Quetico subprovinces. These boundaries might have been brought to the surface by a combination of crustal and magmatic thickening and erosion, followed by both reverse and strike-slip faulting.

Is the KSZ unique? Are there other areas in the Superior Province where field relationships, and/or associated structural, metamorphic or geophysical characteristics are similar to those in the Kapuskasing Structural Zone? We note that 2 other major northeast-trending faults cross the Superior Province: the Gravel River (*see* Williams, this volume) and Miniss River (*see* Breaks, this volume) faults. A brief review of these follows.

The northwestern hanging-wall of the northeast-trending, sinistral, oblique-slip Miniss River Fault, north of Sioux Lookout, brings to the surface granulite-facies supracrustal gneisses (Figure 25.10b). These include anorthositic and mafic plutonic rocks of the Winnipeg River and English River subprovinces, which are some of the highest metamorphic grade rocks outside the KSZ (*see* Beakhouse, this volume; Breaks, this volume; Breaks and Bond 1977; Chipera and Perkins 1988). We consider it highly significant that the Miniss River Fault coincides with the eastern termination of the Winnipeg River Subprovince and corresponds to a significant change in the width of the Wabigoon Subprovince to the south. We propose that these relationships are comparable to those seen farther south in the Wawa Subprovince (*see* Williams, Stott, Heather et al., this volume) which have been interpreted as the product of "indentor" tectonics (Tapponier and Molnar 1976; Ratschbacher, Merle et al. 1991).

A second major fault crossing the Superior Province is the northeast-striking, sinistral-slip Gravel River fault, which is traceable beneath the Moose River Basin, some 100 km north of Hearst. This fault traverses the Quetico, Wabigoon and English River subprovinces (*see* Williams, this volume). The patterns in recently compiled magnetic anomaly data (*see* Map 2586, map case), and Bouguer anomaly data (*see* Map 2594, map case) exhibited by this fault are comparable to those exhibited by the Miniss River Fault and KSZ. The combined field and geophysical data have been interpreted to indicate sinistral kinematics on the fault, and the juxtaposition on the northwest side of highly magnetic and dense rocks against less magnetic, less dense rocks on the southeast side (I.A. Osmani, Ontario Geological Survey, personal communication, 1990). This fault cuts and displaces subprovinces and their boundaries, but unlike the Miniss River Fault, does not mark the termination of a subprovince.

Perhaps the locus of the KSZ was determined in part by an existing weakness, an early terrane boundary between what are now known as the Wawa and Abitibi subprovinces. Such a boundary might have formed during their accretion onto the Quetico and Opatica subprovinces. We note here that although data are sparse, there appears to be a contrast between the metasedimentary Quetico Subprovince and plutonic Opatica Subprovince (*see* Williams, this volume). Evidence in western Quebec (E. Sawyer, Université du Québec à Chicoutimi, personal communication, 1990) indicates that north of the Abitibi Subprovince, the Opatica Subprovince consists predominantly of plutonic igneous rocks and is therefore lithologically analogous to the Winnipeg River Subprovince. The Opatica Subprovince in Ontario is now known to be mainly plutonic in character, not metasedimentary as shown in Figure 1, Map 2545 (map case), although metasedimentary rocks are present. Perhaps the KSZ was located along a zone that marks the western end of the Opatica and coincides with the present boundary between the Abitibi and Wawa subprovinces (Figure 25.10c) that may for a period have developed independently (Jackson and Sutcliffe 1990).

We suggest that these 3 sinistral strike-slip fault zones (KSZ, Miniss River Fault and Gravel River fault) are of similar age, cutting through or forming the terminations of recently formed subprovinces, displacing and enhancing the exposed extent of zones of high metamorphic grade associated with subprovince boundaries by a component of oblique (north over south) slip. The high-grade rocks exposed in these regions provide useful windows into the Archean crust that may be typical of what might underlie large parts of the Superior Province. In the case of the KSZ, the stratigraphic and structural patterns exposed therein are similar to those of high-grade gneiss terranes exposed along subprovince margins (Williams and Breaks 1990). The KSZ was reactivated during the Paleoproterozoic and may then have suffered further uplift.

## ARCHEAN PATTERNS OF MINERALIZATION

A general account of the metallogeny of the Superior Province has been presented by Fyon, Breaks et al. (this volume), and mineralization patterns for each subprovince have been described in preceding chapters. This section briefly reviews the distribution patterns of base metal, platinum group element, rare earth element and gold occurrences in the context of the regional tectonic framework, and highlights the need for a more detailed analysis of the mineral potential of the Superior Province in relation to its lithotectonic assembly, such as that presented by Jackson and Fyon for the Abitibi Subprovince (this volume).

A recent model to explain the distribution and geochemical variation found in volcanogenic massive sulphide (VMS) deposits (Sawkins 1990) proposed that copper-rich deposits are associated with ophiolitic rocks in immature arcs, whereas lead- and zinc-rich VMS deposits are associated with calc-alkalic magmatism in back-arc basins. Many sulphide mineral deposits in the Superior Province are associated with high-silica rhyolites and associated tholeiitic and komatiitic volcanic and intrusive rocks—does this mean that they formed in a back-arc basin environment? Many VMS deposits, such as those at Manitouwadge, Sturgeon Lake, Winston Lake and Kidd Creek, do correlate with the northern boundaries of the present granite-greenstone subprovinces, but the significance of this observation is not understood.

Platinum group elements (PGE) are found in association with intrusive mafic to ultramafic rocks formed towards the end of the Neoarchean (*see* Fyon, Breaks et al., this volume), and tend to be contemporaneous with late granitoid rocks (*see* Sutcliffe et al. 1989). Hybridization of magma, either between successive mafic pulses, or with felsic material, is conducive to sulphide saturation and PGE precipitation.

Rare earth element-bearing pegmatites are typically found along major subprovince boundaries (e.g., English River Subprovince, *see* Breaks, this volume; Uchi Subprovince, *see* Stott and Corfu, this volume), or along faults considered to represent terrane boundaries (*see* Thurston et al., this volume). It appears that the zones separating discrete blocks have acted as loci for the preferential transport of highly differentiated melts.

Kerrich and Wyman (1990) stated that late kinematic, highly evolved subduction-related magmatism during tectonic accretion is the source of gold. Fyon et al. (1989)

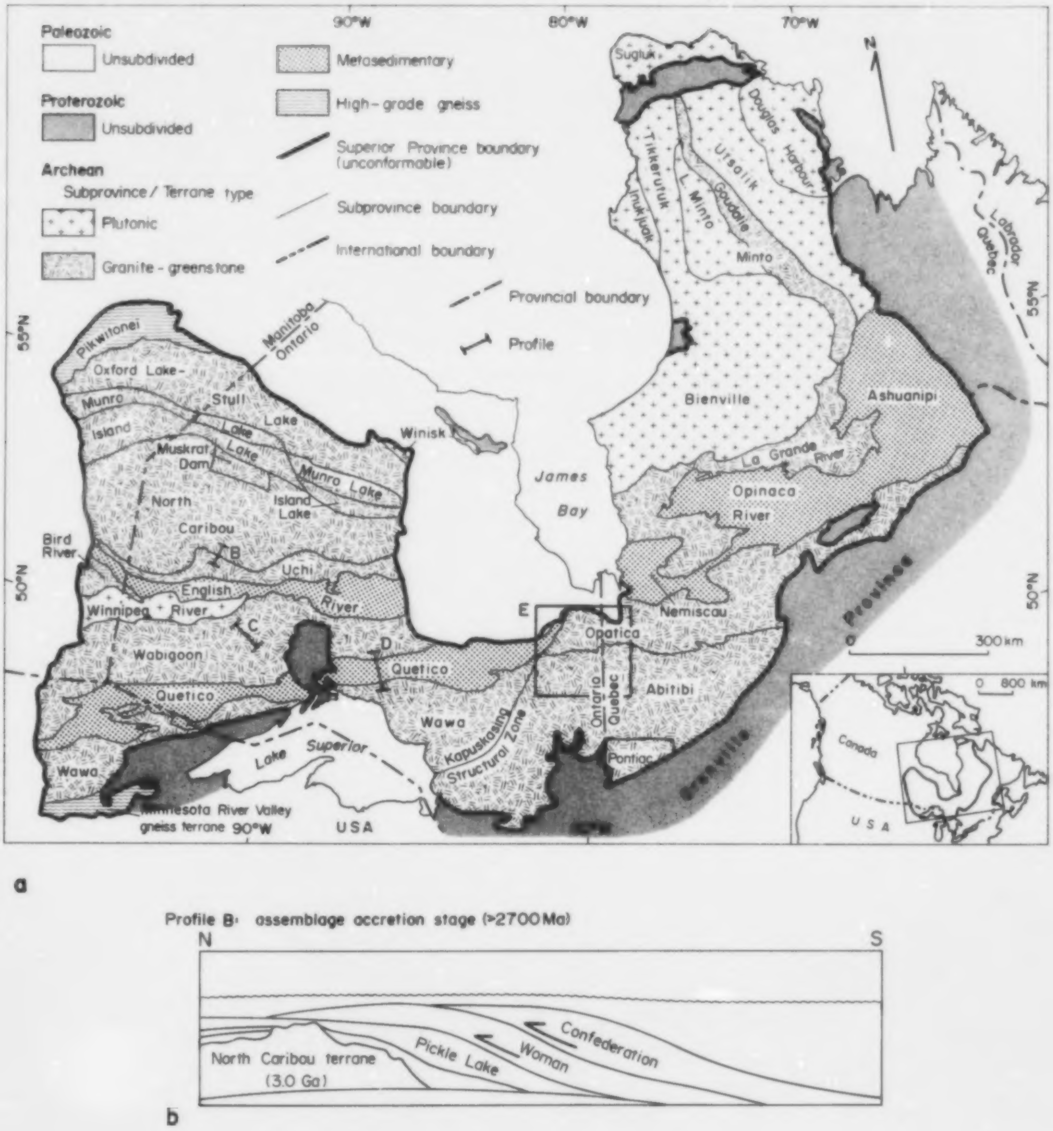

90°W
80°W
70°W
Paleozoic
Unsubdivided
Proterozoic
Unsubdivided
Archean
Subprovince / Terrane type
Plutonic
Granite - greenstone
Metasedimentary
High-grade gneiss
Superior Province boundary (unconformable)
Subprovince boundary
International boundary
Provincial boundary
Profile
N
Sugluk
Tikkerutuk
Inukjuak
L. Minto
Goudalie
Utsalik
Douglas Harbour
Minto
Labrador
Quebec
55°N
Bienville
Ashuanipi
La Grande River
Opinaca River
James Bay
Winisk
Manitoba
Ontario
Pikwitonei
Oxford Lake-
Stull Lake
Munro Lake
Island Lake
Muskrat Dam
North Caribou
Island Lake
Bird River
B
Uchi
English River
Winnipeg River
C
Wabigoon
D
Quetico
Quetico
E
Nemiscau
Opatica
Abitibi
Ontario
Quebec
Kapuskasing Structural Zone
Wawa
Wawa
Lake Superior
Pontiac
Grenville Province
Minnesota River Valley gneiss terrane
USA
0
300 km
0
800 km
Canada
U S A
50°N
a
Profile B: assemblage accretion stage (>2700 Ma)
N
S
Confederation
Woman
Pickle Lake
North Caribou terrane
(3.0 Ga)
b

**Figure 25.10. a)** New map of major tectonic subdivisions of the Superior Province within Ontario, based on data in this volume; data for adjacent areas of Quebec was supplied by J. Percival, Geological Survey of Canada. The tectonic subdivisions are subprovinces, some of which have been tentatively subdivided into terranes. Profiles B through D and inset E are highlighted in successive figures. **b)** Diagram showing the assemblage accretion stage (profile B). Autochthonous supracrustal rocks developed on older sialic crust and allochthonous assemblages were then juxtaposed against the North Caribou terrane, for example, the Woman autochthonous assemblage and the South Uchi parautochthonous assemblage. **c)** Greenstone belt stage (profile C) involving large-scale growth of accreted assemblages through volcanism, tectonism, and varied thickening by the emplacement of tonalite sheets and discordant granitoid plutons. **d)** Subprovince convergence stage (profile D) involving the juxtaposition of accretionary prisms against the collage of greenstone belt assemblages. **e)** Superior Province transpressive deformation stage (inset E) involving late transpressive deformation resulting from province-wide shortening with regional-scale wrench and thrust faulting along subprovince boundaries.

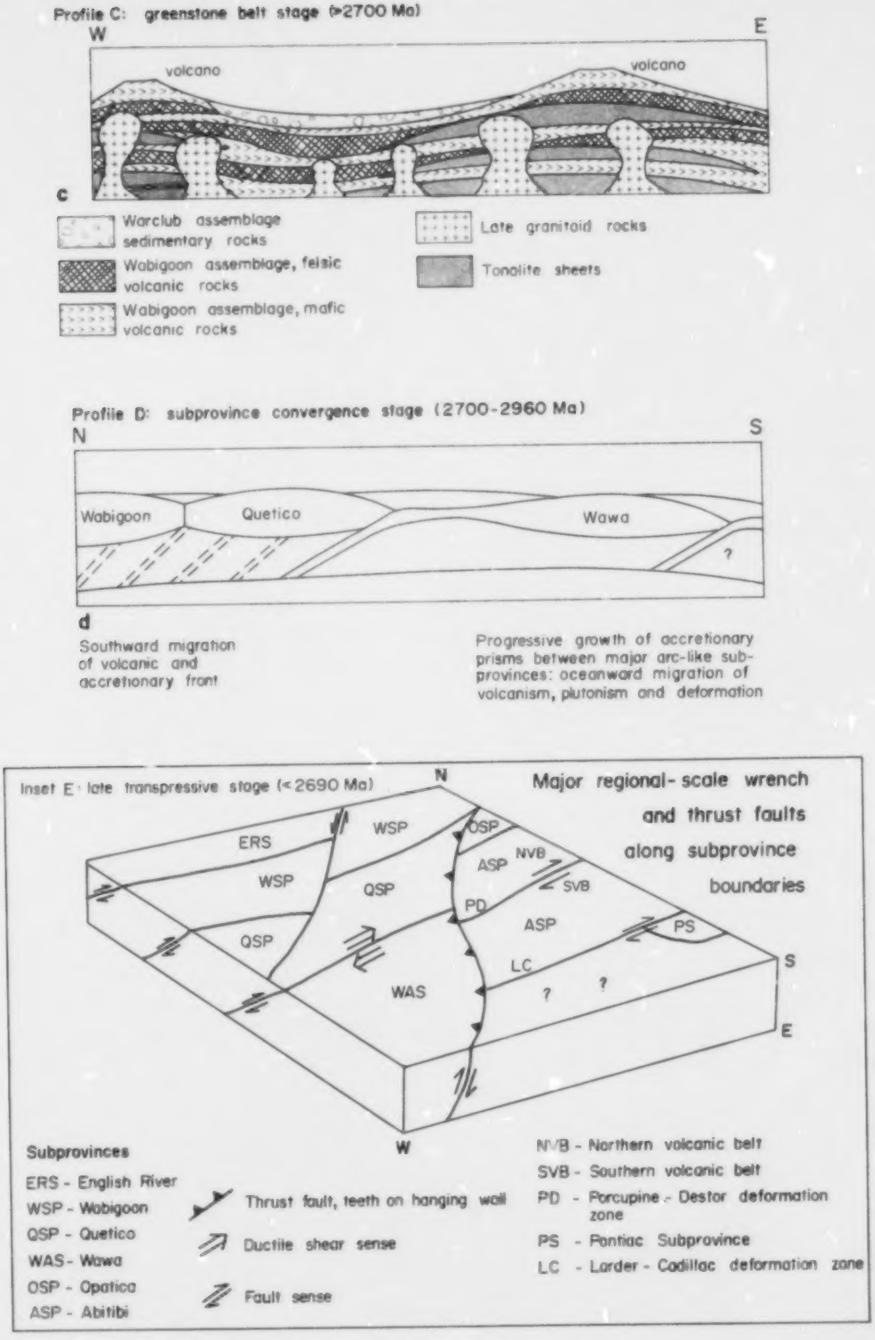

Profile C: greenstone belt stage (>2700 Ma)
W
E
volcano
volcano
c
Warclub assemblage sedimentary rocks
Wabigoon assemblage, felsic volcanic rocks
Wabigoon assemblage, mafic volcanic rocks
Late granitoid rocks
Tonalite sheets
Profile D: subprovince convergence stage (2700-2960 Ma)
N
S
Wabigoon
Quetico
Wawa
?
d
Southward migration of volcanic and accretionary front
Progressive growth of accretionary prisms between major arc-like subprovinces: oceanward migration of volcanism, plutonism and deformation
Inset E: late transpressive stage (< 2690 Ma)
N
Major regional-scale wrench and thrust faults along subprovince boundaries
ERS
WSP
OSP
NVB
ASP
SVB
WSP
QSP
PD
ASP
PS
QSP
LC
S
WAS
?
?
E
W
Subprovinces
ERS - English River
WSP - Wabigoon
QSP - Quetico
WAS - Wawa
OSP - Opatica
ASP - Abitibi
Thrust fault, teeth on hanging wall
Ductile shear sense
Fault sense
NVB - Northern volcanic belt
SVB - Southern volcanic belt
PD - Porcupine - Destor deformation zone
PS - Pontiac Subprovince
LC - Larder - Cadillac deformation zone
e

**Figure 25.10. a)** New map of major tectonic subdivisions of the Superior Province within Ontario, based on data in this volume; data for adjacent areas of Quebec was supplied by J. Percival, Geological Survey of Canada. The tectonic subdivisions are subprovinces, some of which have been tentatively subdivided into terranes. Profiles B through D and inset E are highlighted in successive figures. **b)** Diagram showing the assemblage accretion stage (profile B). Autochthonous supracrustal rocks developed on older sialic crust and allochthonous assemblages were then juxtaposed against the North Caribou terrane, for example, the Woman autochthonous assemblage and the South Uchi parautochthonous assemblage. **c)** Greenstone belt stage (profile C) involving large-scale growth of accreted assemblages through volcanism, tectonism, and varied thickening by the emplacement of tonalite sheets and discordant granitoid plutons. **d)** Subprovince convergence stage (profile D) involving the juxtaposition of accretionary prisms against the collage of greenstone belt assemblages. **e)** Superior Province transpressive deformation stage (inset E) involving late transpressive deformation resulting from province-wide shortening with regional-scale wrench and thrust faulting along subprovince boundaries.

and Powell et al. (1991) favoured gold introduction during post-accretionary crustal differentiation by magmatism, concomitant metamorphism and fluid release. Regardless of this debate concerning the source and timing of gold introduction, many gold-bearing zones overprint and are therefore younger than ductile fabrics related to regional deformation zones. Some gold showings are spatially related to zones of strongly developed mechanical anisotropy defined by major strike-slip shear zones (Osmani and Stott 1988), but some shear zones may have preferentially developed in rocks already altered during preaccretionary volcanic processes.

The relation of gold mineralization to the geological development of the Superior Province has excited much debate (e.g., Colvine 1989; *see* Fyon, Breaks et al., this volume). Despite the common spatial correlation between lode gold and regional-scale zones of high strain, the timing of gold introduction, and its source, are still in question (cf., Powell et al. 1991); thus the tectonic implications of this association are unclear. A link between Archean lode gold deposits and shield cratonization has been furthered by Colvine (1989). Parts of the northern Superior Province were assembled at 2.8 Ga, prior to the Kenoran Orogeny (2.7 Ga) (*see* Thurston et al., this volume; Stott and Corfu, this volume; Fyon, Breaks et al., this volume). The Kenoran Orogeny, which affected the northern and southern flanks of the Superior Province, undoubtedly affected the older central area of the Superior Province (cf., Stott and Corfu, this volume). Isotopic (Gulson et al. 1990) and structural (*see* Stott and Corfu, this volume) data have been used as evidence of a gold mineralizing event prior to 2.8 Ga in, for example, the Red Lake area, but the isotopic data may represent the effects of contamination of gold-mineralizing fluids by 2.8 billion-year-old crust. Thus, there is as yet insufficient evidence of multiple gold mineralizing events in the Superior Province.

## UNIFORMITARIANISM AND THE ARCHEAN

The patterns of tectonic development described and analyzed in this volume, and especially in this chapter, are not radically different from those documented in younger orogens, including those currently active.

We have utilized the purely descriptive terminology recently developed for the examination of modern orogens, but which has been successful in describing ancient ones (e.g., Price and Hatcher 1983). Despite the differences documented in this volume, we have also noted strong parallels between the geology of ancient and modern orogens, parallels that justify our interpretation of the late development of the Superior Province in terms of accreting volcanic arcs, sedimentary prisms and continental fragments (cf., Williams 1990). Despite this similarity, we have emphasised that structural processes governing the scale and kinematics of development of the greenstone collages prior to subprovince-scale interaction have not yet been determined.

The Superior Province shows both similarities and contrasts in age, rock type and structure to other granite-greenstone terranes, such as the Yilgarn and Pilbara cratons, the West African craton, the Slave Province, and the Kaapvaal and Zimbabwe cratons. The similarities include: low- to medium-grade metamorphism, greenstone to granite ratio, and the sequence of lithological units and structural sequence. The differences include: a greater range of ages than in the Superior Province, much older ages in some other cratons, limited evidence in other cratons of the well-defined belt-like greenstone-rich subprovinces alternating with accretionary prisms (cf., Kusky 1989). Recent work in the Abitibi Subprovince (Davis 1991) suggests that sedimentary units (tens of kilometres long and several kilometres thick) on the margins of greenstone assemblages may be relicts equivalent to larger accretionary prisms represented by, for example, the Quetico Subprovince (*see* Williams, this volume). Similar small units in other shields may play comparable roles (cf., Dook 1989) and the Superior Province may only be unique for the size of its sedimentary subprovinces.

## DISCUSSION

In this section, we present a number of somewhat speculative discussions of aspects of the accretionary model.

### Significance of the Arc Accretion Model

The model proposed here for the evolution of the Superior Province is uniformitarian. Notwithstanding, for example, the secular physical changes associated with a cooling Earth, including the paucity of komatiites in recent orogens, we have observed that over the last 3 billion years of the Earth's history, there has been surprisingly little change in the style, scale and causes for geological processes and the resultant rocks.

### Continuity and Shape of Subprovinces

In Figure 25.7, we proposed a scheme for the sequential accretion of tectonic units to produce the Superior Province. In Figure 25.10a, we have revised the subprovince map of Ontario to include the revisions proposed in this volume; our interpretation indicates that all subprovinces are not continuous along the tectonic "grain" of the Superior Province. For example, the Winnipeg River Subprovince appears to die out eastward; similarly, the plutonic part of the Opatica Subprovince, located predominantly in Quebec, dies out westward. Based on a proposal by Stott (1986), we suggest that these geometric relationships and the senses of shear along the resultant zones of slip between subprovinces (Figure 25.10b), may represent an example of "indentor" tectonics (Tapponnier and Molnar 1976; Ratschbacher et al. 1991; Ratschbacher, Merle et al. 1991), whereby triangular promontories on the northern edges of subprovinces impinge on already accreted masses. In some cases, the promontory appears to cut or coincide with the lateral termination of a subprovince (e.g., the Miniss River Fault forms the southeast boundary of the Winnipeg River Subprovince against the Wabigoon Subprovince; *see* Figure 25.10b).

Similarly, the northeast Abitibi–Opatica subprovince boundary perhaps forms the western termination of the Opatica Subprovince. However, in the case of Wawa–Quetico relationships, the promontory in the Shebandowan region is flanked similarly by the Quetico Subprovince (*see* Williams, Stott, Heather et al., this volume).

## Significance of Accretionary Prisms in the Archean

Sedimentary rock-dominated subprovinces compare closely with modern accretionary prisms. The volume of metasedimentary rock in the Quetico and English River subprovinces is testament to the amount of quickly erodible material shed from the adjacent volcanic arcs. The rate of accumulation of tectonized sediment in modern prisms varies, and depends on plate motion rate, the angle of plate subduction and the efficiency of underplating mechanisms (Abbott and Hoffman 1984). For example, estimates derived from data in Hibbard and Karig (1990) and Nishi (1988) suggest that it took about 10 million years to develop the Shimanto accretionary prism (southwest Japan), which is about 100 km wide, but consists of shallowly dipping thrust panels with a structural thickness of about 20 km. Similarly, the Kodiak prism, part of the Chugach Terrane (Sample and Moore 1987), developed a width of 70 km in 12 million years . The approximately 70 km wide Quetico Subprovince also developed within about 10 to 15 million years, but this age range is only weakly constrained. The determination of better constrained rates of sediment accumulation in the Quetico Subprovince await more age determinations. Despite the undoubted secular change in Earth's heat production since the Archean, it is remarkable how similar the Quetico prism is in size and rate of formation to modern analogues.

Rocks accreted at the leading edge of an accretionary prism young progressively trenchward (e.g., Shimanto Belt or accretionary prism, Japan, Taira et al. 1988), a direction opposite to way-up directions. Lacking the necessary paleontological or geochronological control, such age distribution cannot be established within the sedimentary rock-dominated accretionary prisms such as the Quetico and English River subprovinces. In the granite-greenstone subprovinces, a progressive southward younging of accreted assemblages towards the accretionary prism has been noted (e.g., Uchi Subprovince, *see* Stott and Corfu, this volume; Wabigoon Subprovince, *see* Blackburn et al., this volume). Subprovince-scale structural asymmetry and the younging of assemblages leads us to infer that the polarity of subduction-trench systems for both the Quetico and English River prisms was north dipping and south migrating.

In modern tectonic environments, the cessation of subduction-related volcanism, accretionary prism development and the consequent preservation of the prism between complex aggregations of volcanic assemblages is caused by the underthrusting of the leading edge of the prism by an isostatically buoyant volcanic arc. We suggest that an analogous situation developed at the end of the Archean, when the accretionary prisms now recognized as the Quetico and English River subprovinces grew on the flanks of the Wabigoon and Uchi subprovinces, respectively, and became trapped against incoming terranes and superterranes, now recognized as the Wawa and Wabigoon–Winnipeg River subprovinces, respectively.

# FUTURE RESEARCH

The following section is a list of questions that we believe should form the basis of continued research. At first sight some may appear academic, but they have profound implications with regard to the ore-bearing potential of rocks in Ontario, and elsewhere. We consider that future research should focus on the following outstanding problems:

1. Multidisciplinary studies are required to define, document and test the nature of the various postulated tectonic settings outlined in this chapter.
2. What proportion of the sedimentary rock in greenstone belts is the result of assemblage juxtaposition? Do some represent "thrust-tip shedding" in the form of small accretionary prisms?
3. Much more detailed analysis of the granitoid rocks between greenstone belts is required to discriminate their tectonic settings.
4. We need to understand the controls on hydrothermal alteration patterns. Synvolcanic regional hydrothermal alteration patterns favourable to volcanogenic massive sulphide mineralization may be related to extensional settings in volcanic arcs and back-arc basins. Other hydrothermal alteration patterns formed during orogenesis may be related to midcrustal thickening due to magmatism and/or crustal stacking accompanied by magmatism. What are the implications of such crustal thickening with regard to gold mineralization? Do regions that have undergone more than 1 thickening event have a higher gold potential?
5. Is the volume of greenstone volcanic rocks greater in Archean accretionary complexes than in modern orogens (Isozaki et al. 1990), and, if so, how does this relate to the production and chemistry of magmatism in greenstone belts?
6. Was a greater volume of oceanic mafic rocks accreted during the Archean than in modern orogens? If so, what are the implications of this difference? Did the thermal properties of rapidly generated and rapidly spreading Archean oceanic crust make it less amenable to subduction, and therefore more prone to, accretion?
7. Where and how do we look for more components predating the Meso- to Neoarchean accretion?
8. What was the relationship of the Minnesota River Valley gneiss terrane to the Wawa and Abitibi subprovinces?
9. Why do old assemblages appear to congregate around batholiths?
10. Detailed paleomagnetic studies of assemblages or groups of assemblages are required to test the thesis of large relative motions.

11. Are "thick-skinned" hinterland thrust faults produced during the progressive accretionary additions from the south?
12. Indentor tectonics are responsible for characteristic sets of structures (e.g., Ratschbacher, Frisch et al. 1991; Ratschbacher, Merle et al. 1991); can these be documented in Archean terranes?

# PART 2: PROTEROZOIC GEOLOGY OF THE GREAT LAKES REGION: A DEFORMED CRATON MARGIN AND THE MIDCONTINENT RIFT

**R.H. Sutcliffe and G. Bennett**

## INTRODUCTION

In this section, the Proterozoic geologic and tectonic evolution of Ontario, exclusive of the Grenville Province, is summarized. The material presented within the 5 separate chapters in this volume on the Proterozoic of Ontario which elucidate large-scale relationships is integrated, and a brief summary of the rocks of the Trans–Hudson Orogen is provided. The latter is poorly exposed in Ontario and, hence, is not described in a separate chapter. This section will be more condensed than the Superior Province tectonic summary (*see* Part 1: Revolution in the Superior Province) because Proterozoic rocks occupy a relatively small area in Ontario, and the tectonic framework for the Proterozoic has been recently reviewed (Larue 1983; Hoffman 1988, 1989).

The North American craton consists of several Archean provinces (e.g., Superior, Wyoming, Hearne), separated by linear Proterozoic belts (Figure 25.11) which in Ontario are:

1. the 1.9 to 1.8 billion-year-old Trans–Hudson Orogen, which separates the Superior and Hearne provinces and underlies a small part of northwestern Ontario, largely beneath the Paleozoic and Mesozoic Hudson Bay Basin
2. the 2.4 to 2.2 billion-year-old Penokean Orogen, which separates the Superior and Yavapai–Mazatzal provinces and is exposed along the shores of Lake Huron and Lake Superior
3. the 1.1 billion-year-old Midcontinent Rift, which is filled with volcanic and sedimentary rocks up to 30 km thick, making it one of the largest plateau basalt systems on Earth and is exposed in the vicinity of Lake Superior

The Proterozoic orogens are characterized by sequences, which are at scales and patterns that are readily relatable to modern tectonic analogues. In this chapter, we will summarize the major events in the Trans–Hudson and Penokean orogens and comment upon controls on Proterozoic events within the Superior Province.

## TRANS–HUDSON OROGEN

The Trans–Hudson Orogen is the collision zone between the Superior Province and the Hearne and Wyoming provinces. Hoffman (1989) divided the orogen into segments: the Hudson Bay segment and northern Quebec segments to the north and northeast, and the Manitoba–Saskatchewan and Dakota segments to the northwest. Here, we deal mainly with the Hudson Bay segment and to a lesser extent, the Manitoba–Saskatchewan segment, parts of which underlie Ontario. The former is composed (Figure 25.12), from southeast to northwest, of the Nastapoka homocline, the Belcher fold belt, the Winisk trough, the Severn arc and the Central Hudson Bay hinterland. The Severn arc is a possible magmatic arc in central Hudson Bay postulated on geophysical evidence (Roksandic 1987; Hoffman 1989). The hinterland lies north of the Severn arc in Hudson Bay.

The Nastapoka homocline is composed of the autochthonous Richmond Gulf Group, a basal shelf sequence of fluvial arkose, mafic metavolcanic rocks and fine-grained clastic and carbonate sedimentary rocks surmounted by a foredeep sequence of basal quartz arenite and succeeded by iron formation, mafic volcanic rocks, euxinic shale and turbidite. Diagenetic apatite gives an age of 2.02 Ga for the basal sandstone (Chandler 1988). The sequence is largely derived from the west. The marine strata forming the upper part of the unit are exposed in the Belcher fold belt to the west. Rocks correlatable with the Nastapoka homocline occur in Ontario along the flanks of the Cape Henrietta Maria Arch (Hoffman 1989), where Bostock (1971) has mapped a sequence of undeformed carbonate and clastic sedimentary rocks in the Sutton Ridges area. The Sutton Ridges units are in turn correlated (Weber 1990) with rocks of the Fox River Belt in Manitoba.

In the Sutton Ridges area (Figure 25.13), the homocline comprises siliceous carbonate of the Nowashe Formation, disconformably overlain by iron formation and clastic rocks of the Sutton Ridges Formation (Bostock 1971). The Nowashe Formation is a 75 m thick stromatolite-bearing dolostone with minor siliceous calcareous argillite, limestone and dolomitic limestone. The Sutton Ridges Formation, 120 m thick, comprises a basal chert breccia conglomerate succeeded upward by wacke and silicate- and carbonate-facies iron formation. The rocks have undergone limited contact metamorphism related to 1884 million-year-old diabase sills and dikes of the Molson swarm.

In Ontario, the Belcher fold belt northeast of the Sutton Ridges is not exposed. This unit is composed of the folded equivalents of the Nastapoka homocline (Hoffman 1989; Roksandic 1987). Based on seismic (Roksandic 1987) and aeromagnetic data (*see* Map 2584, map case), the fold belt occurs northeast of the Sutton Ridges, buried beneath the rocks of the Hudson Bay Basin (*see* Johnson et al., this volume).

Northeast (in Ontario) of the Belcher fold belt is the Winisk trough, bounded by normal faults and containing well-stratified Proterozoic sedimentary rocks which are inferred from geophysical data (Roksandic 1987) to be at least 6 km thick. The Winisk trough lies mainly in the Northwest Territories beneath the waters of Hudson Bay, with only a minor part underlying far northwestern Ontario. The Winisk trough is considered to be an episutural basin between the Chipewayan arc and the ensialic sedimentary domain of the Nastapoka homocline (Hoffman 1989). The Chipewayan arc or Severn arc, north of the Winisk trough, is composed of volcanic and plutonic rocks exposed largely in Manitoba.

The Manitoba–Saskatchewan segment of the orogen is divided by Hoffman (1989) into a southeastern foreland belt, an internal zone of juvenile Proterozoic crust, an Andean-type batholith and a hinterland on the northwest side (*see* Figure 25.12). The foreland belt is composed of the Fox River Belt, which in Manitoba (Scoates 1981) consists of a north-dipping thrusted sequence of clastic metasedimentary rocks and tholeiitic to komatiitic flows and sills, including the 1.88 billion-year-old Fox River sill (Scoates 1981). The Fox River Belt is succeeded to the north by a sequence of interarc turbidites that are correlative (Hoffman 1989) with the Kisseynew gneisses of Manitoba. The Fox River Belt is bounded on the south successively by Archean

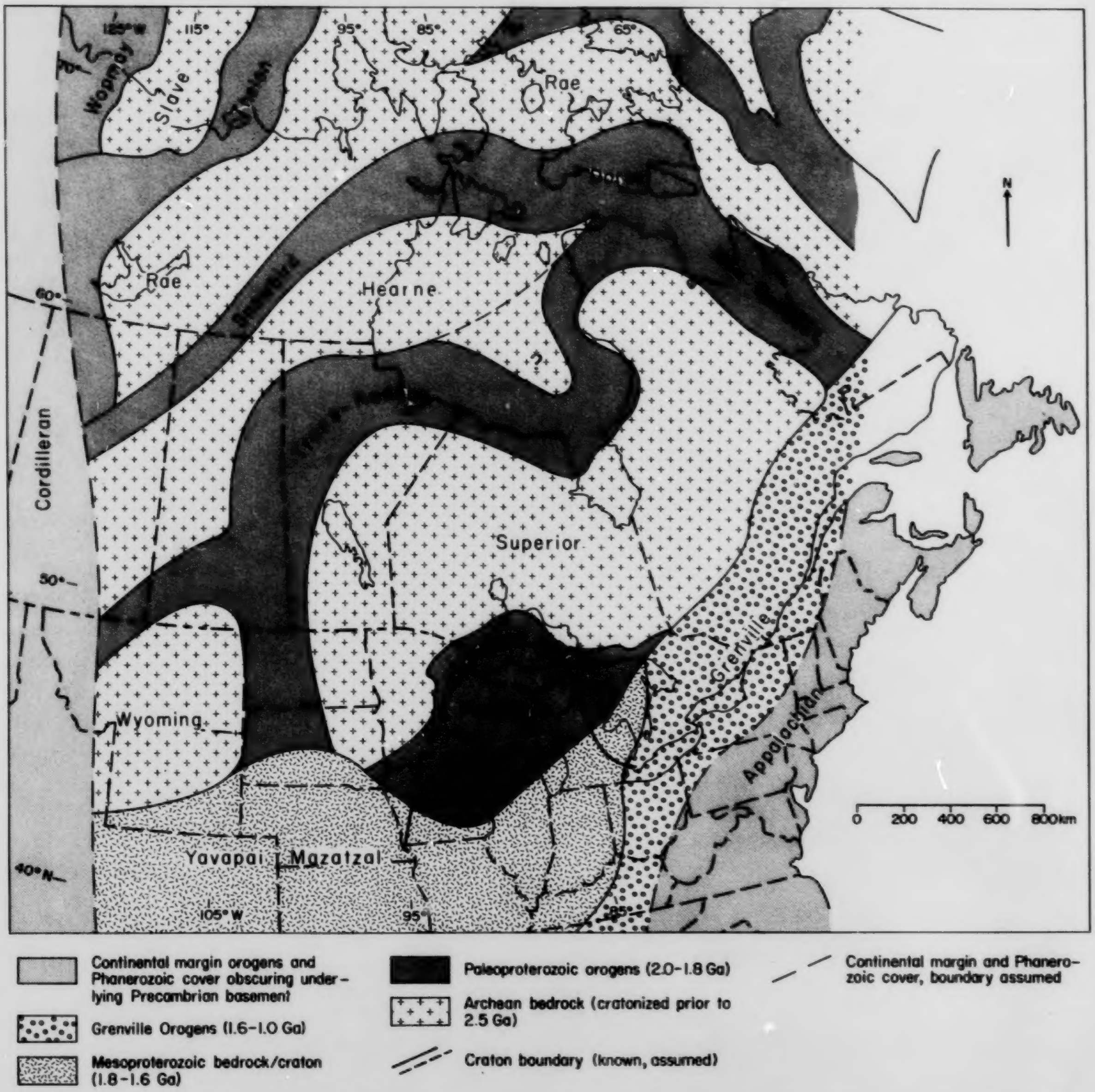

125°W
115°
95°
85°
65°
70°
Wopmay
Slave
Thelon
Rae
Rae
Snowbird
Hearne
60°
?
Trans-Hudson
Cordilleran
Superior
50°
Wyoming
Grenville
Appalachian
Yavapai
Mazatzal
40°N
105°W
95°
85°
N
0 200 400 600 800km
Continental margin orogens and Phanerozoic cover obscuring underlying Precambrian basement
Grenville Orogens (1.6–1.0 Ga)
Mesoproterozoic bedrock/craton (1.8–1.6 Ga)
Paleoproterozoic orogens (2.0–1.8 Ga)
Archean bedrock (cratonized prior to 2.5 Ga)
Craton boundary (known, assumed)
Continental margin and Phanerozoic cover, boundary assumed

**Figure 25.11.** The craton of Laurentia showing Archean cratons bordered by Proterozoic orogens. The terminology is taken from Hoffman (1989).

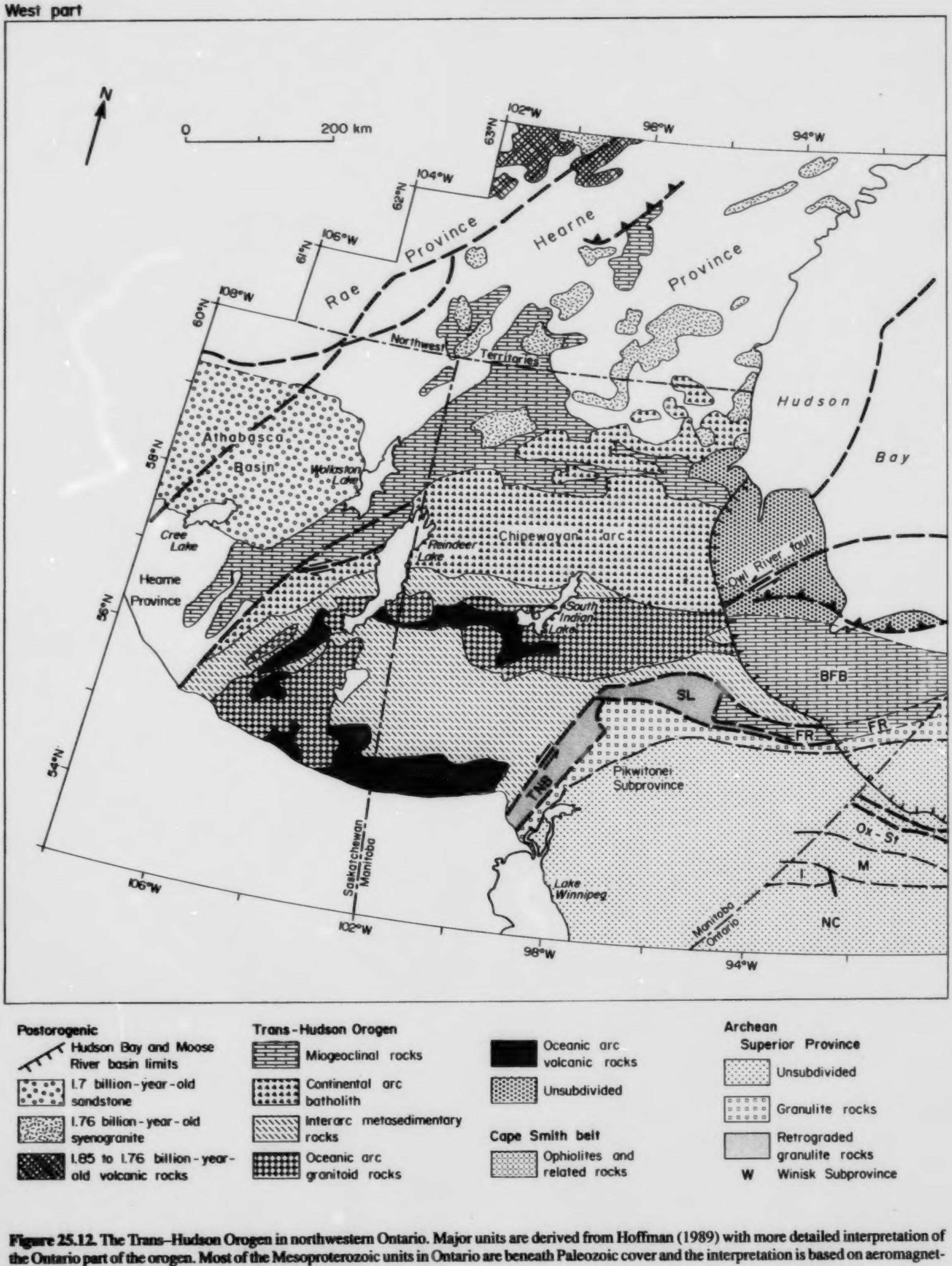

West part
N
0 200 km
102°W
98°W
94°W
104°W
106°W
108°W
63°N
62°N
61°N
60°N
58°N
56°N
54°N
Rae Province
Hearne Province
Northwest Territories
Hudson Bay
Athabasca Basin
Wollaston Lake
Cree Lake
Hearne Province
Reindeer Lake
Chipewayan arc
South Indian Lake
Owl River fault
BFB
SL
FR
FR
TNB
Pikwitonei Subprovince
Ox - St
M
I
NC
Lake Winnipeg
Saskatchewan Manitoba
Manitoba Ontario
106°W
102°W
98°W
94°W
Postorogenic
Hudson Bay and Moose River basin limits
1.7 billion-year-old sandstone
1.76 billion-year-old syenogranite
1.85 to 1.76 billion-year-old volcanic rocks
Trans-Hudson Orogen
Miogeoclinal rocks
Continental arc batholith
Interarc metasedimentary rocks
Oceanic arc granitoid rocks
Oceanic arc volcanic rocks
Unsubdivided
Cape Smith belt
Ophiolites and related rocks
Archean
Superior Province
Unsubdivided
Granulite rocks
Retrograded granulite rocks
W Winisk Subprovince

**Figure 25.12.** The Trans–Hudson Orogen in northwestern Ontario. Major units are derived from Hoffman (1989) with more detailed interpretation of the Ontario part of the orogen. Most of the Mesoproterozoic units in Ontario are beneath Paleozoic cover and the interpretation is based on aeromagnetic and gravity data (Maps 2584 and 2592, respectively).

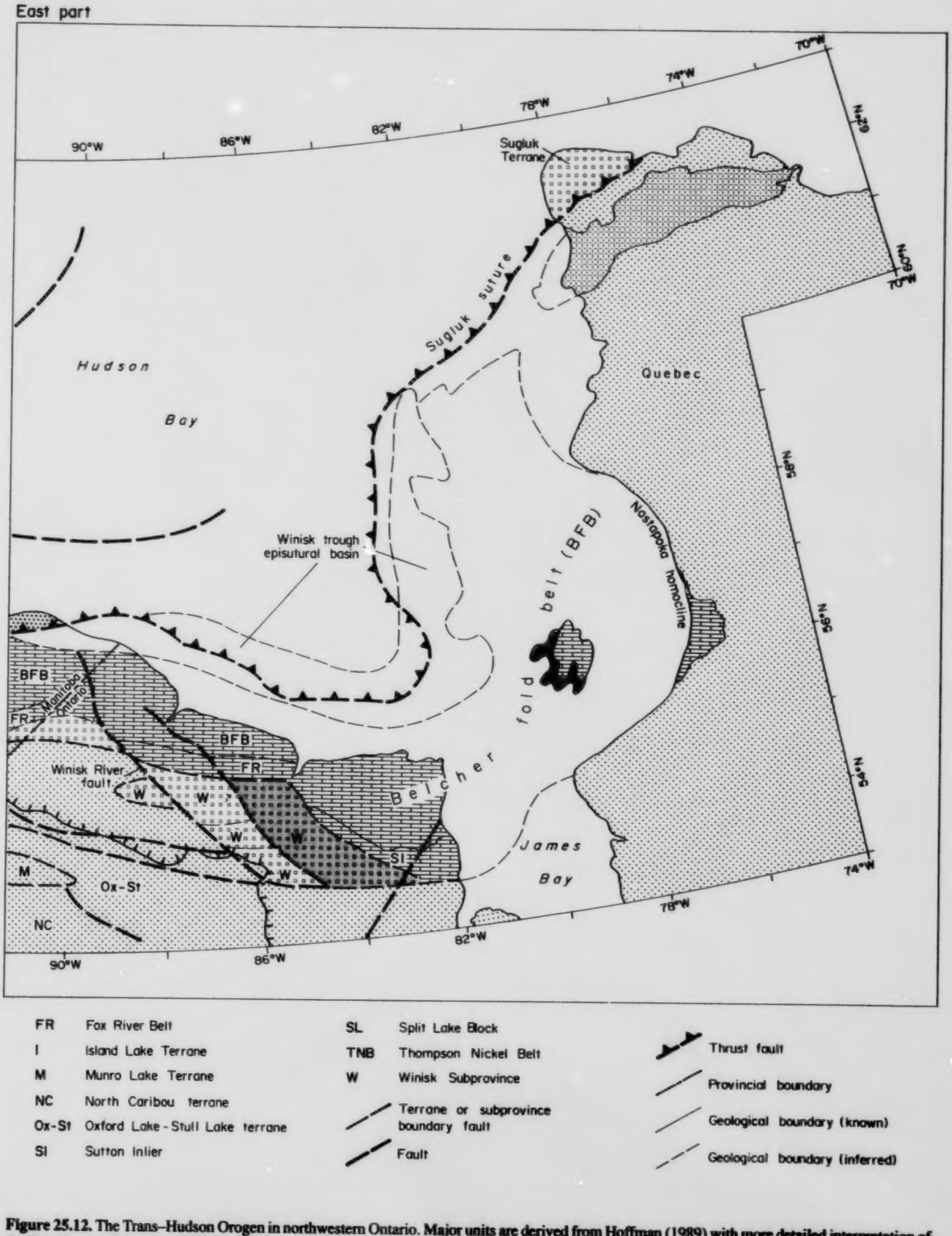

East part
70°W
74°W
78°W
82°W
86°W
90°W
62°N
60°N
58°N
56°N
54°N
Sugluk Terrane
Sugluk suture
Hudson Bay
Quebec
Winisk trough episutural basin
Nastapoka homocline
Belcher fold belt (BFB)
BFB
Manitoba Ontario
FR
Winisk River fault
W
SI
M
Ox-St
NC
James Bay
FR Fox River Belt
I Island Lake Terrane
M Munro Lake Terrane
NC North Caribou terrane
Ox-St Oxford Lake-Stull Lake terrane
SI Sutton Inlier
SL Split Lake Block
TNB Thompson Nickel Belt
W Winisk Subprovince
Terrane or subprovince boundary fault
Fault
Thrust fault
Provincial boundary
Geological boundary (known)
Geological boundary (inferred)

Figure 25.12. The Trans–Hudson Orogen in northwestern Ontario. Major units are derived from Hoffman (1989) with more detailed interpretation of the Ontario part of the orogen. Most of the Mesoproterozoic units in Ontario are beneath Paleozoic cover and the interpretation is based on aeromagnetic and gravity data (Maps 2584 and 2592, respectively).

granulite-facies rocks of the Pikwitonei Subprovince and the Thompson Nickel Belt, the latter being composed of high metamorphic rank Archean gneissic rocks and a Paleoproterozoic foredeep sequence (Bleeker 1990) and ultramafic rocks hosting nickel deposits near Thompson, Manitoba. The Thompson Nickel Belt was believed to extend into Ontario in the vicinity of the Winisk River fault (Ayres et al. 1969). In the Superior Province, the pyroxene-bearing granitic rocks adjacent to the Winisk River fault (Riley *in* Thurston et al. 1979) and the general younging westward and northward of K-Ar ages toward the margin of the province represent the effects of the Trans–Hudson Orogen on the Superior Province of Ontario.

The Fox River Belt can be traced aeromagnetically across northern Manitoba into Ontario (*see* Map 2584, map case), where it is dextrally offset along the Winisk River fault (Thurston et al. 1979; *see* Figure 25.12). The east-trending deformed rocks of the Fox River Belt pass abruptly southward into east-northeast-trending highly magnetic rocks of the Winisk Subprovince, which are here interpreted in part as granulite facies equivalents of the Pikwitonei Subprovince in Manitoba. This interpretation is based on the proximity of the Winisk Subprovince to the Fox River Belt rocks and the similarity of the aeromagnetic expression of both the Pikwitonei and parts of the Winisk subprovinces. The contact between the Winisk Subprovince rocks and the west-northwest-trending undeformed rocks of the Sutton Ridges area is unexposed, but the lack of rotation of the foliation of the Winisk Subprovince units near the contact suggests that rocks of the Sutton Ridges area were

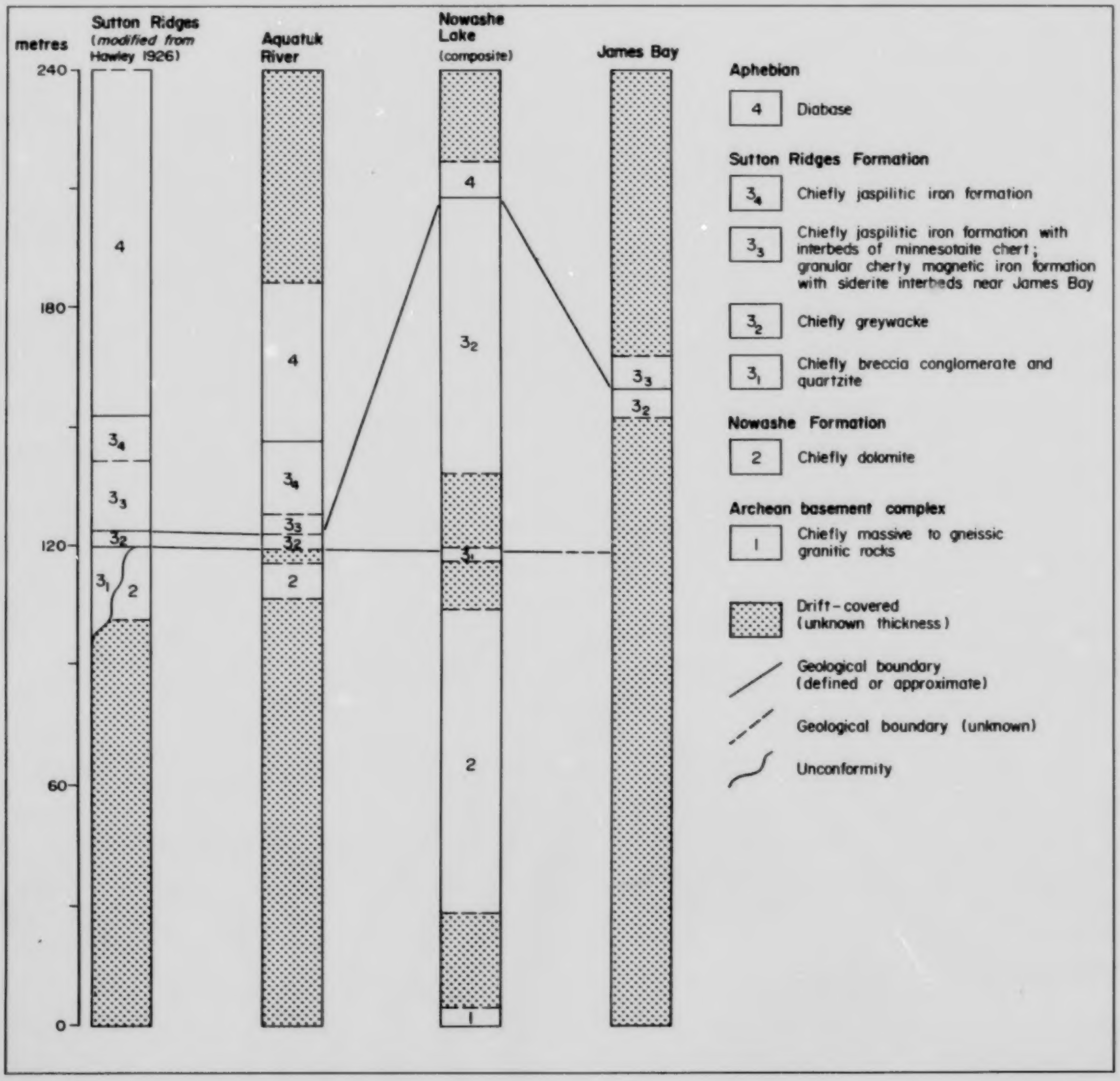

metres
240
180
120
60
0
Sutton Ridges (modified from Hawley 1926)
Aquatuk River
Nowashe Lake (composite)
James Bay
4
3₄
3₃
3₂
3₁
2
1
Aphebian
4 Diabase
Sutton Ridges Formation
3₄ Chiefly jaspilitic iron formation
3₃ Chiefly jaspilitic iron formation with interbeds of minnesotaite chert; granular cherty magnetic iron formation with siderite interbeds near James Bay
3₂ Chiefly greywacke
3₁ Chiefly breccia conglomerate and quartzite
Nowashe Formation
2 Chiefly dolomite
Archean basement complex
1 Chiefly massive to gneissic granitic rocks
Drift-covered (unknown thickness)
Geological boundary (defined or approximate)
Geological boundary (unknown)
Unconformity

**Figure 25.13.** Stratigraphic sections of Mesoproterozoic rocks of the Sutton Ridges area (*modified from* Bostock 1971).

probably deposited unconformably upon rocks of the Winisk Subprovince, consistent with the geochronologic evidence (Riley *in* Thurston et al. 1979).

The northwest-southeast-directed compression associated with the Trans–Hudson Orogeny may have produced or reactivated intracratonic faults within the Superior Province, for example, the Kapuskasing Structural Zone (Percival 1989b; Figure 25.14). The principal evidence for the *circa* 1.8 billion-year-old activity of the Kapuskasing Structural Zone are the *circa* 1.8 billion-year-old isotopic ages documented in its vicinity. The Trans–Hudson Orogen may also have been the cause of the development of 1.8 billion-year-old low temperature hydrothermal minerals associated with the Eye–Dashwa Pluton in the Wabigoon Subprovince (Kamineni et al. 1990). Such secondary minerals are likely to be widespread, but little work has been done on the ages of late low temperature minerals in the Superior Province.

## PENOKEAN OROGEN

The Penokean Orogen lies along the southern margin of the Superior Province and consists of a *circa* 2.4 billion-year-old deformed continental margin prism in the north, succeeded southward by a sequence of oceanic arcs, the Wisconsin magmatic terranes (Figure 25.15). The Niagara Suture Zone (Sims et al. 1989) forms the southern limit of the Superior Province. South of the suture, the Wisconsin magmatic terranes consist of the northern Pembine–Wausau Terrane and the southern Marshfield terrane juxtaposed along the Eau Pleine suture (Sims et al. 1989). The Niagara Suture Zone (1.86 to 1.835 Ga) is characterized by serpentinite, sheeted dikes, plagiorhyolite and tholeiitic basalts. The Pembine–Wausau Terrane is a sequence of deformed volcanic and related sedimentary rocks intruded by 1.89 to 1.88 billion-year-old granitic to gabbroic plutons. Nd isotopic data (Barovich et al. 1989) indicate that the terrane is

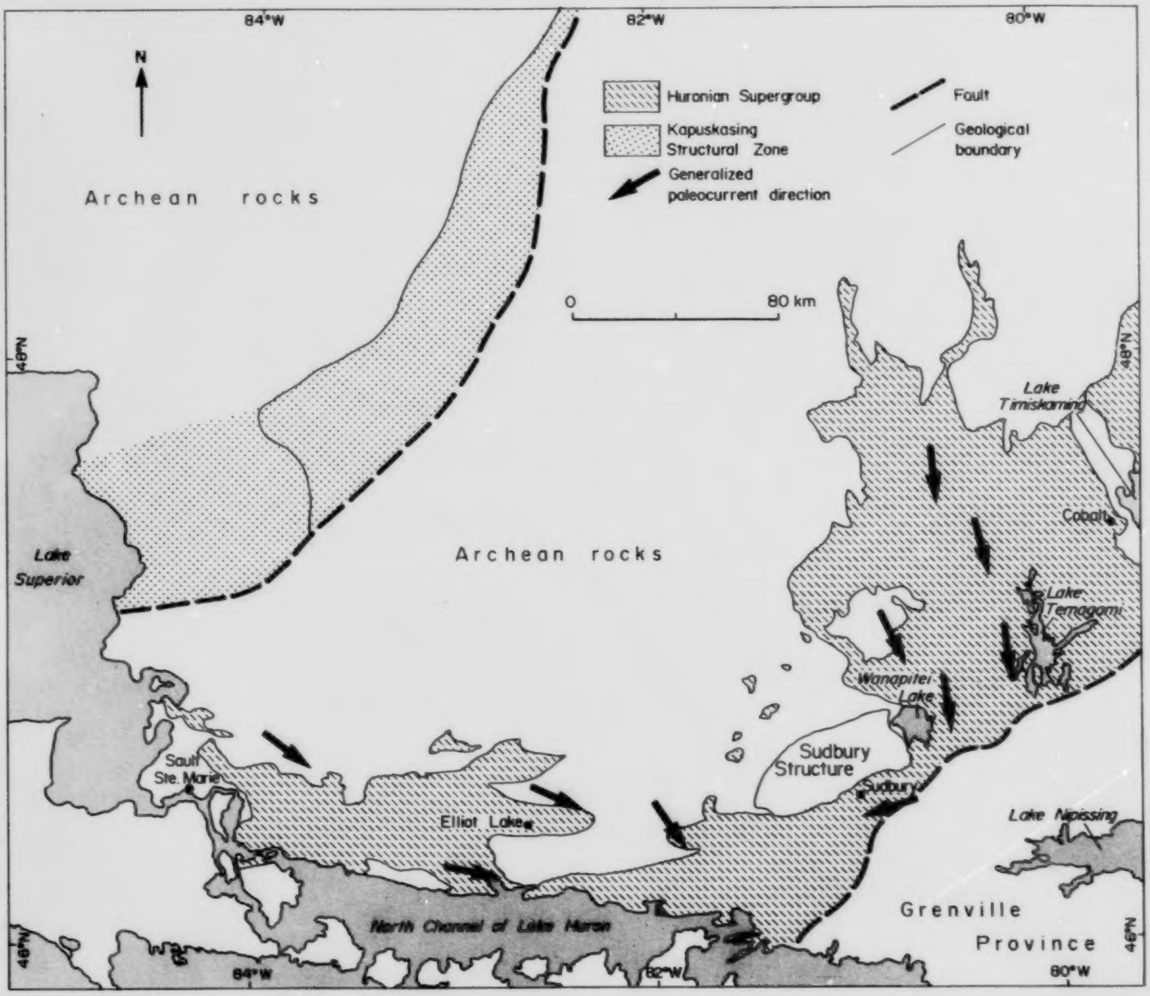

84°W
82°W
80°W
N
Huronian Supergroup
Kapuskasing Structural Zone
Generalized paleocurrent direction
Fault
Geological boundary
Archean rocks
0
80 km
48°N
48°N
Lake Timiskaming
Cobalt
Lake Superior
Archean rocks
Lake Temagami
Wanapitei Lake
Sudbury Structure
Sudbury
Sault Ste. Marie
Elliot Lake
Lake Nipissing
Grenville Province
North Channel of Lake Huron
46°N
46°N
84°W
82°W
80°W

**Figure 25.14.** The Kapuskasing Structural Zone uplift as a possible source for detritus in the upper part of the Huronian Supergroup based on paleocurrent data (*see* Bennett et al., this volume).

derived from Paleoproterozoic or juvenile crust with a minor Archean component. The terrane is inferred to be an island arc and/or back-arc basin that accreted to the Superior Province.

The Marshfield terrane consists of a 3.0 to 2.5 billion-year-old Archean basement overlain by terrigenous sedimentary rocks and 1.86 billion-year-old volcanic rocks cut by 1.89 to 1.84 billion-year-old tonalitic, mafic and granitic

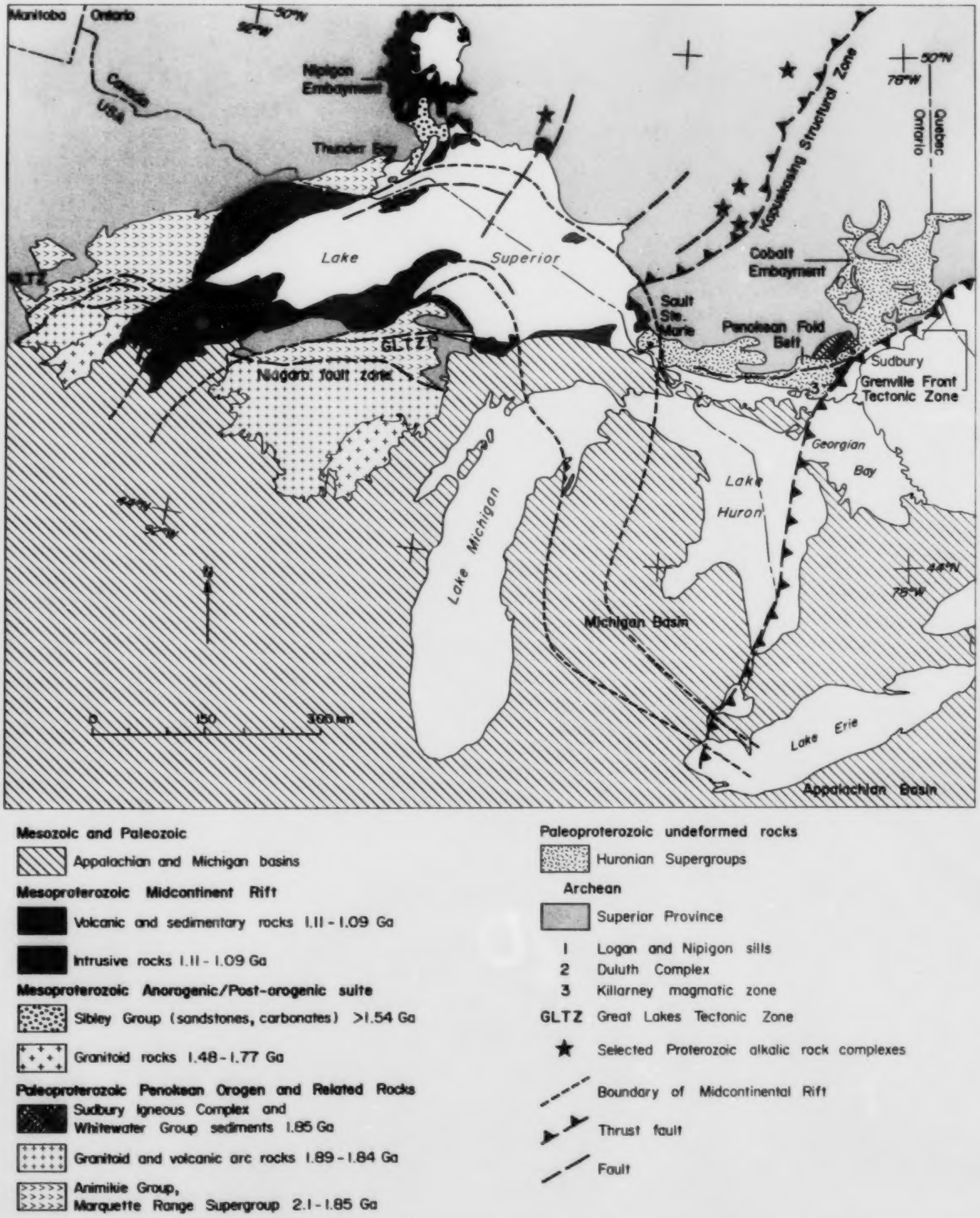

Manitoba
Ontario
50°N
92°W
Canada
USA
Nipigon Embayment
Thunder Bay
Lake Superior
GLTZ
Niagara fault zone
Kapuskasing Structural Zone
50°N
78°W
Quebec
Ontario
Cobalt Embayment
Sault Ste. Marie
Penokean Fold Belt
Sudbury
Grenville Front Tectonic Zone
3
Georgian Bay
Lake Huron
Lake Michigan
Michigan Basin
44°N
92°W
44°N
78°W
N
0
150
300 km
Lake Erie
Appalachian Basin
Mesozoic and Paleozoic
Appalachian and Michigan basins
Mesoproterozoic Midcontinent Rift
Volcanic and sedimentary rocks 1.11 - 1.09 Ga
Intrusive rocks 1.11 - 1.09 Ga
Mesoproterozoic Anorogenic/Post-orogenic suite
Sibley Group (sandstones, carbonates) >1.54 Ga
Granitoid rocks 1.48 - 1.77 Ga
Paleoproterozoic Penokean Orogen and Related Rocks
Sudbury Igneous Complex and Whitewater Group sediments 1.85 Ga
Granitoid and volcanic arc rocks 1.89 - 1.84 Ga
Animikie Group, Marquette Range Supergroup 2.1 - 1.85 Ga
Paleoproterozoic undeformed rocks
Huronian Supergroups
Archean
Superior Province
1 Logan and Nipigon sills
2 Duluth Complex
3 Killarney magmatic zone
GLTZ Great Lakes Tectonic Zone
★ Selected Proterozoic alkalic rock complexes
Boundary of Midcontinental Rift
Thrust fault
Fault

**Figure 25.15.** Sketch map of the major units of the Great Lakes region (*modified from* Sutcliffe, this volume).

plutons. This terrane is interpreted by Hoffman (1989) to be an Archean continental fragment along the south margin of a 1.8 billion-year-old arc, the Pembine–Wausau Terrane.

The Archean is overlain in Minnesota, Wisconsin and Ontario by a deformed continental margin sedimentary prism consisting of a lower, rifted passive-margin sequence (Morey 1983), overstepped to the north by a synorogenic foredeep sequence (Hoffman 1987). The former includes the Marquette Range Supergroup in Michigan and Wisconsin (Cannon and Gair 1970), and the broadly correlative Huronian Supergroup in Ontario. The Huronian Supergroup is composed of 4 unconformity bounded groups of siliciclastic rocks which successively overlap northward across the Superior Province. The age of the unit is constrained by the 2450 million year age of volcanic rocks within the lowermost of the 4 groups and the 2219 million year age of crosscutting Nipissing diabase bodies. The equivalents to the Huronian Supergroup in the USA are overstepped northward by a foredeep sequence. Rocks of the foredeep sequence pass upward from a basal transgressive marine quartzite through ironstone, euxinic shale, tholeiitic basalt and wacke turbidite. Southeast of the Mesoproterozoic Midcontinental Rift, the foredeep sequence is known as the Marquette Range Supergroup (Morey 1983), whereas northwest of the rift, the sequence is split by the Duluth Gabbro into the Mesabi and Cuyuna iron ranges in Minnesota and north of the gabbro, the Animikie Group. The Animikie Group progresses from basal conglomerate, carbonate and iron formation of the Gunflint Formation through to argillite of the Rove Formation. The rock sequence represents the products of deposition in an epicontinental basin, which deepened as a result of the north-verging fold-thrust belt emanating from the Penokean Orogen to the south.

The southern margin of the Superior Province is the site of the 1.85 billion-year-old Sudbury Igneous Complex, an elliptical norite-granophyre lopolith emplaced into brecciated country rock. The complex is overlain by a basin filled by the Whitewater Group comprising breccias of enigmatic origin, euxinic pelite and turbidite.

In the Mesoproterozoic, the southern margin of the Superior Province was subject to anorogenic plutonism consisting (Sutcliffe and Greenwood 1982) of predominantly subvolcanic 1.54 billion-year-old (Davis and Sutcliffe 1985) granitic rocks and minor fragmental pyroclastic rocks preserved beneath 1.1 billion-year-old Logan sills in the Lake Nipigon area. In addition, felsic clasts preserved in the *circa* 1.1 billion-year-old Osler Formation sedimentary rocks suggest felsic magmatism at 1.6 and 1.7 Ga as well (*see* Sutcliffe, this volume). The 1.7 to 1.6 billion-year-old clasts are coeval with the accretion and deformation of the Mazatzal-Labradorian Orogen. The Penokean Orogen is cut by 1.77 to 1.74 billion-year-old granite-rhyolite bodies in the USA; felsic rocks in the Killarney Magmatic Belt (van Breemen and Davidson 1988a) may be similar (cf., Easton, this volume). The Penokean Orogen and adjacent areas of the Superior Province are overlain by erosional remnants of the Sioux and Baraboo quartz arenites which predate the Midcontinent Rift. These are not exposed in Ontario. The Superior Province is overlain by the *circa* 1.3 billion-year-old (Mesoproterozoic) Sibley Group, undeformed epicontinental clastic rocks consisting of quartz arenite, argillaceous dolostone and mudstone, interpreted to have been deposited in a playa lake. The sequence is preserved in a north-trending graben east and south of Lake Nipigon.

The North American craton was split by the Midcontinent Rift at about 1.1 Ga. The northern end of the rift extends beneath Lake Superior and eastward through Michigan, crossing the Grenville Front Tectonic Zone south of Lake St. Clair in southwestern Ontario beneath the Paleozoic rocks of the Michigan Basin (*see* Figure 25.15). Rocks of the Keweenawan Supergroup were deposited within and marginal to the rift in a series of linked half grabens of alternating structural polarity. Keweenawan rocks consist of a lower 1109 to 1094 million-year-old (Davis and Sutcliffe 1985; Davis and Paces 1990) sequence of subaerial tholeiitic basalt and subordinate felsic units, overlain by alluvial-fluvial sedimentary rocks of the Oronto Group in turn overlain by an upper sedimentary unit, the fluvial sedimentary rocks of the Jacobsville Formation. In Ontario, the lower part of the Keweenawan Supergroup is composed of: 1) the Osler Formation east of Thunder Bay; 2) the Mamainse Point Formation and the Michipicoten Island Formation on Michipicoten Island; and 3) the Mamainse Point Formation in the Mamainse Point area west of Sault Ste. Marie. The upper part of the supergroup is a sequence of red beds of the Oronto Group deposited during and after subsidence following the 1094 million-year-old volcanism in the rift, unconformably overlain by the quartzose clastic sedimentary rocks of the Jacobsville Formation. Exposures of the upper part of the supergroup occur in the eastern Lake Superior area. Seismic reflection data (Cannon et al. 1989) indicate the rift is up to 30 km deep, making the rocks of the lower Keweenawan Supergroup one of the largest plateau basalt accumulations on Earth. Rifting at 1.1 Ga was contemporaneous with major thrusting of the Grenville Province onto the craton (Indares and Martignole 1989; *see* Easton, this volume). Some suggest this synchroneity implies a linkage between crustal thickening in the orogen and thinning in the foreland brought about by northwest-directed shortening (Hoffman 1989). However, the Midcontinent Rift is markedly arcuate with parts trending northeast, parallel to the Grenville Front and therefore perpendicular to principal shortening. The process of rifting for the northeast-trending part of the Midcontinent Rift is not readily explained by a mechanism involving the Grenville Orogen unless periods of extension occurred during the orogeny.

## PROTEROZOIC MINERALIZATION

Proterozic mineral deposits reflect a variety of tectonic environments and genetic processes. Mineral deposits of this era are reviewed in the metallogeny chapter (*see* Fyon, Bennett et al., this volume).

In the Paleoproterozoic, the foredeep environment of the Animikie Group contains the major oxide facies iron deposits of the Minnesota and Michigan iron ranges, but the correlative Gunflint Formation contains no currently economical deposits. Aboriginals traded flint from the Gunflint Formation for native copper from deposits on the Upper Peninsula of Michigan. The paleoplacer uranium deposits at Elliot Lake were produced by the erosion of the Archean rocks to the north, and the incorporation of uranium into the basal part of a continental margin sedimentary prism represented by the Huronian Supergroup (*see* Bennett et al., this volume; *see also* Fyon, Bennett et al., this volume). The origin of the copper-nickel-platinum group element deposits in the Sudbury Igneous Complex is enigmatic and may be due to meteoritic impact or to an endogenic process. The Mesoproterozoic in Ontario has potential for several deposit types associated with the felsic anorogenic magmatism; however, no deposits are known. Mafic magmatism associated with the Keweenawan Supergroup includes copper-nickel-platinum group element deposits associated with the Crystal Lake Gabbro in the Thunder Bay area and the Port Coldwell alkalic complex. Veins and breccia bodies associated with Keweenawan Supergroup rocks contain minor copper mineralization and the silver deposits of the Thunder Bay area (*see* Fyon, Bennett et al., this volume). The breccia bodies are also the source of amethyst mineralization near Thunder Bay (*see* Fyon, Bennett et al., this volume). Silver-sulpharsenide vein deposits, associated with the Huronian Supergroup in the Cobalt area and the Keweenawan Supergroup in the Thunder Bay area, were mined beginning in the 19th century. Both camps were instrumental in developing the present resource-related pattern of settlement of the province.

# PART 3: MESOPROTEROZOIC EVOLUTION OF THE SOUTHEAST MARGIN OF LAURENTIA

**R.M. Easton**

## INTRODUCTION

The Grenville Province is a complex orogenic belt of *circa* 1.1 billion year age that truncates several older geologic provinces, including: the Superior and Southern provinces, the Killarney Magmatic Belt, the Eastern Granite–Rhyolite Province and the Midcontinent Rift. Recorded within the Grenville Province is the tectonic evolution of the southeast margin of Laurentia during the Mesoproterozoic. Just as the Acadian and Taconic orogenies in the Paleozoic Appalachian Orogen overprinted pre-existing structural trends and geologic provinces, so has the Grenville Orogeny overprinted the Archean and Paleoproterozoic geologic provinces of Laurentia. In this section, the geologic and tectonic evolution of the Grenville Province and Laurentia is summarized.

Supracrustal rocks of the Huronian Supergroup affected by the 1.8 billion-year-old Penokean Orogeny are separated from the Grenville Front by plutonic and minor volcanic rocks of the Killarney Magmatic Belt (van Breemen and Davidson 1988a; Figure 25.16). Intrusive rocks of the Killarney Magmatic Belt cut the Huronian Supergroup (Davidson 1986b) and are 1750 to 1730 and 1470 to 1450 million years old (van Breemen and Davidson 1988a). Southeast of the Grenville Front, these rocks were affected by the Grenville Orogeny. The Killarney Magmatic Belt is interpreted to be an extension of one of several Mesoproterozoic orogenic belts underlying much of the North American midcontinent (Easton 1986; van Breemen and Davidson 1988a).

Except near Manitoulin Island, basement rocks in southwestern Ontario have been affected by a *circa* 1140 to 1070 million-year-old orogenic event, the "Grenville Orogeny." The Grenville Front marks the northwesterly limit of penetrative Grenvillian metamorphism and deformation and truncates structural trends in the adjacent structural provinces (*see* Figure 25.16), although locally some of these rocks can be traced across the Grenville Front. It is now generally accepted that this orogenic event (or events) involved northwest-directed thrusting and imbrication of the entire crust, presumably as a result of a terminal collision at about 1140 Ma with a continental landmass (cf., Moore 1986; Rivers et al. 1989).

## Tectonic Framework of the Grenville Province

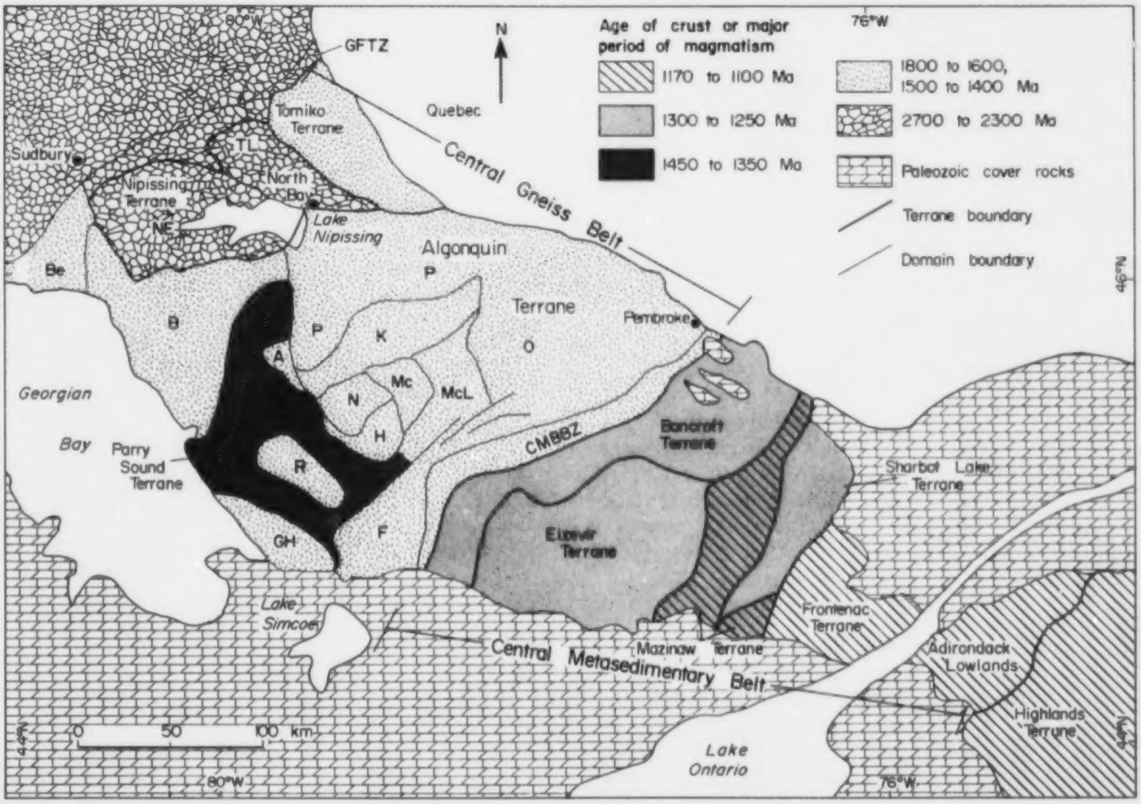
There are 2 main divisions of the Grenville Province in Ontario; the Central Gneiss Belt and the Central Metasedimentary Belt (Wynne-Edwards 1972; *see* Figure 25.16). The Central Gneiss Belt within 20 to 30 km of the Grenville Front is termed the Grenville Front Tectonic Zone (e.g., Wynne-Edwards 1972) and consists of deformed and metamorphosed rocks derived from the Southern and Superior provinces and the Killarney Magmatic Belt.

### CENTRAL GNEISS BELT

The Central Gneiss Belt consists mainly of upper amphibolite- and local granulite-facies quartzofeldspathic gneisses chiefly of igneous origin with subordinate paragneiss. Distinctive lithotectonic terranes, and subordinate domains, identified within the Central Gneiss Belt (*see* Easton, this volume; *see* Figure 25.16) are distinguished by differences in rock types, internal structure, metamorphic grade, geologic history and geophysical signature: they are bounded by kilometre-scale deformation zones (Davidson 1984a; Rivers et al. 1989). Similar terranes and domains can be identified in the subsurface of southwestern Ontario (Carter and Easton 1990) where they influenced Paleozoic deposition (*see* Easton, this volume).

As shown in Figure 25.16, rocks of 3 main ages are present in the Central Gneiss Belt (Easton 1986; Dickin et al. 1990): north of the French River, reworked Archean and Paleoproterozoic gneisses of the Nipissing Terrane are intruded by several ages of Mesoproterozoic largely granitic and monzonitic intrusions. The bulk of the Central Gneiss Belt (Algonquin and Tomiko terranes) consists of 1.8 to 1.6 billion-year-old gneisses intruded by 1.5 to 1.4 billion-year-old granitic and monzonitic plutons that may represent an extension of the Eastern Granite–Rhyolite Province across the Grenville Front. The Tomiko Terrane may consist entirely of Mesoproterozoic rocks that are allochthonous with respect to the adjacent Nipissing Terrane (*see* Easton, this volume). The Algonquin Terrane probably represents a section of 1.8 to 1.6 billion-year-old crust extensively injected by granitic magmas of the Eastern Granite–Rhyolite Province. Although imbricated by later thrusting, the Algonquin Terrane is probably parautochthonous (Rivers et al. 1989).

Overlying and southeast of these predominantly parautochthonous terranes, throughout the Grenville Province in Canada, are several allochthonous terranes having no obvious relationships with the terranes they overlie (Rivers et al. 1989). Within the Central Gneiss Belt in Ontario only one such allochthonous terrane, the Parry Sound Terrane, has been identified. It (*see* Figure 25.16) consists of mafic to intermediate rocks extracted from the mantle at about 1.45 to 1.35 Ga (Dickin and McNutt 1989a) and in contrast to the adjacent Algonquin Terrane, it consists entirely of juvenile crust. The Fishog Domain (Algonquin Terrane) may also be an allochthonous terrane, as it is characterized by an intense magnetic high and a distinctive plutonic suite (Easton 1990). No isotopic ages are available from the


80°W
GFTZ
N
Quebec
Tomiko Terrane
TL
Sudbury
Nipissing Terrane
NE
North Bay
Lake Nipissing
Central Gneiss Belt
Algonquin Terrane
Be
B
P
K
O
A
Mc
McL
N
H
R
GH
F
Pembroke
Georgian Bay
Parry Sound Terrane
CMBBZ
Bancroft Terrane
Elzevir Terrane
Sharbot Lake Terrane
Frontenac Terrane
Mazinaw Terrane
Lake Simcoe
Central Metasedimentary Belt
Adirondack Lowlands
Highlands Terrane
Lake Ontario
0 50 100 km
44°N
46°N
76°W
Age of crust or major period of magmatism
▨ 1170 to 1100 Ma
1300 to 1250 Ma
■ 1450 to 1350 Ma
1800 to 1600, 1500 to 1400 Ma
2700 to 2300 Ma
Paleozoic cover rocks
Terrane boundary
Domain boundary

**Abbreviations**

| | |
|---|---|
| A | Ahmic Domain |
| B | Britt Domain |
| Be | Beaverstone Domain |
| CMBBZ | Central Metasedimentary Belt Boundary Zone |
| F | Fishog Domain |
| GFTZ | Grenville Front Tectonic Zone |
| GH | Go Home Domain |
| H | Huntsville Domain |
| K | Kiosk Domain |
| Mc | McCraney Domain |
| McL | McClintock Domain |
| MR | Moon River Domain |
| N | Novar Domain |
| NE | Nepewassi Domain |
| O | Opeongo Domain |
| P | Powassan Domain |
| PS | Parry Sound Domain |
| R | Rosseau Domain |
| S | Seguin Domain |
| TL | Tilden Lake Domain |

**Figure 25.16.** Lithotectonic terranes, domains and crustal ages within the Central Gneiss Belt and the Central Metasedimentary Belt in Ontario, based on Davidson (1984b, 1986a), Easton (1986, 1988), Moore (1982), Culshaw et al. (1983) and Dickin and McNutt (1989a, 1989b). Algonquin Terrane consists of the Ahmic, Britt, Fishog, Go Home, Huntsville, Kiosk, McCraney, McClintock, Novar, Opeongo, Powassan and Rosseau domains.

Fishog Domain, consequently it is not known if it is allochthonous (i.e., a terrane), or if it is part of the Algonquin Terrane (i.e., a domain).

## CENTRAL METASEDIMENTARY BELT

The Central Metasedimentary Belt is a major Mesoproterozoic accumulation of supracrustal rocks invaded by compositionally diverse plutons of various ages. The entire succession has been metamorphosed at grades varying from greenschist to granulite facies. The Central Metasedimentary Belt can be divided into several lithotectonic terranes (*see* Figure 25.16).

Two discrete regions exist within the Central Metasedimentary Belt:

1. a superterrane composed of the Bancroft, Elzevir, Mazinaw and Sharbot Lake terranes, including the classical Grenville Supergroup, characterized by volcanism and sedimentation between 1300 and 1250 Ma, followed by plutonism and metamorphism at 1250 to 1240 Ma and at 1140 to 1070 Ma
2. the Frontenac Terrane, which lacks volcanic rocks and contains a sequence of marbles, quartzites and quartzofeldspathic gneisses that were intruded by plutonic rocks and subjected to metamorphism at roughly 1170 to 1160 Ma

The Frontenac Terrane is preserved at granulite facies, in contrast to the lower grade superterrane to the northwest and the amphibolite facies Adirondack Lowlands to the south.

Mineral deposits of the Elzevir Terrane consist of sulphides (pyrite and chalcopyrite predominate), carbonate-hosted zinc, magnetite skarns, gold in quartz-carbonate vein systems commonly associated with shear zones, and a variety of industrial minerals. Mineral deposits in the Frontenac Terrane include graphite and other industrial minerals. A series of 1090 to 1070 million-year-old syenite intrusions occurs in all terranes of the Central Metasedimentary Belt (Corriveau et al. 1990) and may date the time when all the terranes began to behave as a single tectonic entity.

The Central Metasedimentary Belt Boundary Zone marks the boundary between the Central Gneiss Belt and the allochthonous terranes of the Central Metasedimentary Belt, which have been thrust onto the Central Gneiss Belt from the southeast. The boundary zone (*see* Figure 25.16) is a major shear zone several kilometres wide, characterized by strongly deformed rocks with northeasterly trending, moderately to shallowly southeast-dipping tectonic layering and southeast-plunging mineral lineations. It is subparallel with the Grenville Front Tectonic Zone in Ontario (*see* Figure 25.16). Both zones have pronounced seismic expressions (Green et al. 1988; White et al. 1991) and extend to the Moho. It is not known if, and to what extent, the Central Gneiss Belt served as basement to the supracrustal rocks of the Central Metasedimentary Belt, although there is growing evidence that the Central Gneiss Belt and the Central Metasedimentary Belt are allochthonous and only became attached to one another during the Grenville Orogeny (cf., Davidson 1991).

## CONCEPTS OF OROGENY IN THE GRENVILLE PROVINCE

Regional geochronologic studies in the early 1960s indicated that the Grenville Province was characterized by *circa* 1000 to 950 million- year-old K-Ar biotite ages (Cumming et al. 1955; Stockwell 1964). These ages were attributed to postmetamorphic cooling coincident with the "Grenville Orogeny." It was initially believed that most structural and metamorphic features within the Grenville Province resulted from the *circa* 1.0 billion year Grenville Orogeny. With the advent of Rb-Sr and U-Pb geochronologic studies, however, it became clear that older rocks were present within the Grenville Orogen. This led some workers, such as Wynne- Edwards (1972) to suggest that much of the Grenville Province consisted of reworked Archean and Paleoproterozoic crust, and that little juvenile crust existed in the region. Moore and Thompson (1980) and Baer (1981), on the basis of stratigraphic and geochronologic evidence, suggested that 2 orogenies were present within the Central Metasedimentary Belt, an older (ca. 1240 to 1180 Ma) "Elzevirian Orogeny" and a younger (ca. 1100 Ma) "Ottawan Orogeny", the latter being equivalent to the "Grenville Orogeny." Consensus in the past has been that the younger orogeny was the more extensive, although this view has been questioned recently (Lumbers et al. 1990; *see* Easton, this volume).

## TECTONIC SUMMARY

Much of the history of the various terranes of the Grenville Province is summarized in Figures 25.17 and 25.18, which are generalized time-space charts. Figure 25.19 is a cladogram outlining how the disparate pieces of the Grenville Province may have been assembled. The Grenville Orogen is a complex thrust stack assembled over a considerable time interval. Terranes now adjacent to one another may have been separated by great distances in the past. Figures 25.17 to 25.19 illustrate the complex nature of the Grenville Province in Ontario, and the difficulty in developing comprehensive models. As described below, major intervals of geologic activity coincide with events occurring elsewhere in Laurentia.

### Pre-1780 Million-Year-Old Activity

Rocks older than 1780 Ma, restricted to Segments 2 and 3 of the Grenville Front Tectonic Zone and the Nipissing Terrane (*see* Easton, this volume), are metamorphosed equivalents of Southern and Superior province rocks. The age of "Grenville" metamorphism of these rocks is unknown, although K-Ar ages in the region are in the 1000 to 900 million year range and U-Pb zircon ages of pegmatites in the area, including lower intercept ages, lie between 1000 to 990 Ma. The intensity of "Grenville" metamorphism in this region is not clearly known. There is no evidence of any events unique to the Grenville Province between 2650 and 1780 Ma.

## 1780 to 1680 Million-Year-Old Activity

A major period of crustal development in the Central Gneiss Belt, particularly in the Algonquin Terrane, is indicated by Nd-model ages (Dickin and McNutt 1989a; Dickin et al. 1990) from paragneisses. In addition, younger orthogneisses in the Algonquin Terrane yield *circa* 1.8 billion-year-old model ages, suggesting derivation from mantle or lower crust of this age. Nd-model ages on paragneisses in the Frontenac Terrane (Marcantonio et al. 1990) suggest that the Frontenac Terrane may be a sliver of Central Gneiss Belt-type crust that was rifted from, and then accreted to, Laurentia during the Elzevirian Orogeny.

In the Algonquin Terrane, this interval is first marked by sedimentation (ca. 1780 to 1700 Ma) on a basement of unknown character of probable Penokean age. In the Tomiko Terrane, metasedimentary rocks were deposited at roughly 1687 Ma (Krogh 1989), perhaps a continuation of the *circa* 1780 to 1700 million-year-old sedimentation in the Algonquin Terrane. In the southern Nipissing Terrane, supracrustal rocks (French River area) were deposited at *circa* 1744±11 Ma (Krogh 1989), possibly on Archean or Paleoproterozoic basement. In the Tomiko Terrane, there is no indication of Archean zircons within the metasedimentary rocks (Krogh 1989), indicating isolation from Archean source regions. This contrasts with the Nipissing Terrane, which contains Archean zircons (Krogh 1989), indicating provenance with an Archean component. Thus, although the Nipissing Terrane may be parautochthonous, the Algonquin and Tomiko terranes may have been allochthonous to the Superior craton during their initial development.

In the Killarney area, *circa* 1740 million-year-old volcanism and plutonism may have occurred in a continental arc. In the Nipissing Terrane, the Sturgeon and West Bay granites were emplaced at *circa* 1740 Ma (Lumbers and Vertolli 1991), perhaps coincident with Killarney magmatism. In the Algonquin Terrane, sedimentation was followed by granite intrusion and migmatization (ca. 1680 Ma, Culshaw et al. 1991), the latter probably associated with regional metamorphism. The 1680 million-year-old orogeny included formation of a broad region of juvenile continental crust, further stabilized during a 1450 to 1420 million-year-old anorthosite-mangerite-charnockite-granite (AMCG) magmatic event.

## 1450 to 1420 Million-Year-Old Activity

Major AMCG-type felsic magmatism (Lumbers and Vertolli 1991) occurred throughout the Central Gneiss Belt at *circa* 1450 to 1420 Ma, most notably in the Algonquin Terrane. Regional metamorphism, locally reaching granulite-facies conditions, extended from the Grenville Front Tectonic Zone near Killarney (Krogh 1989), through the Britt Domain (Culshaw et al. 1991). Throughout much of the Central Gneiss Belt, however, this early granulite metamorphism has been overprinted and largely obliterated (Culshaw et al. 1991). Evidence of *circa* 1420 million-year-old magmatism (McLelland et al. 1988) is found also within the Frontenac Terrane.

Nd-model ages (Dickin and McNutt 1990) suggest that the Parry Sound Terrane consists predominantly of material derived from the mantle at *circa* 1400 Ma. The Parry Sound Terrane may represent the basal part of an island-arc complex (Easton 1989; Culshaw et al. 1990). Lithologically, parts of the Grimsthorpe Domain in the Central Metasedimentary Belt are similar to the Parry Sound Terrane, but, at present there is no geochronologic or lithologic link between the two. Rocks similar to those of the Parry Sound Terrane occur in the subsurface of southwestern Ontario in the Cambridge Domain (Figure 25.20).

## 1350 to 1320 Million-Year-Old Activity

A second pulse of magmatism, dominated by tonalitic magmas, occurred in the southern Central Gneiss Belt, and farther to the southeast at *circa* 1350 to 1320 Ma. Anorthositic magmas in the Parry Sound Terrane also crystallized at this time. Rocks of this age may be basement to parts of the Central Metasedimentary Belt, however, such basement would be isotopically transparent.

## 1300 to 1240 Million-Year-Old Activity

The initial time of development of the Central Metasedimentary Belt is difficult to establish, especially the relationship between the different domains within the Elzevir Terrane. The situation is analogous to that observed in Archean greenstone belts of the Superior Province (*see* Part 1: Revolution in the Superior Province) where it is now recognized that greenstone belts are composed of several volcano-sedimentary assemblages ("domains"), generally in tectonic contact. Although different tectonic settings can be determined for these assemblages, their assembly into belts is still problematic (*see* Part 1: Revolution of the Superior Province).

Within the Elzevir Terrane, the Grimsthorpe Domain, composed of tholeiitic mafic volcanic and volcaniclastic rocks (ca. 1300 to 1280 Ma) intruded by tonalitic (ca. 1270 Ma) plutonic complexes, probably developed in a marginal basin or back-arc setting. The slightly younger (ca. 1260 to 1250 Ma), more diverse mafic-felsic volcanic and siliceous clastic and carbonate supracrustal assemblage of the Belmont Domain, Harvey–Cardiff Arch and the Sharbot Lake Terrane is underlain by an older mafic volcanic sequence possibly constructed on a Grimsthorpe Domain-type basement. This younger supracrustal sequence also represents a marginal basin setting, perhaps formed adjacent to a craton, but more likely adjacent to a "continental fragment." The Bancroft Terrane could represent the platform of this "continental" fragment, which had only recently formed. The younger supracrustal sequence is not intruded by typical arc plutons, but by diorite-gabbro (Lavant Suite) and alaskitic granites (Methuen Suite; both ca. 1245 to 1240 million years in age), the latter showing geochemical characteristics typical of anorogenic magmat-

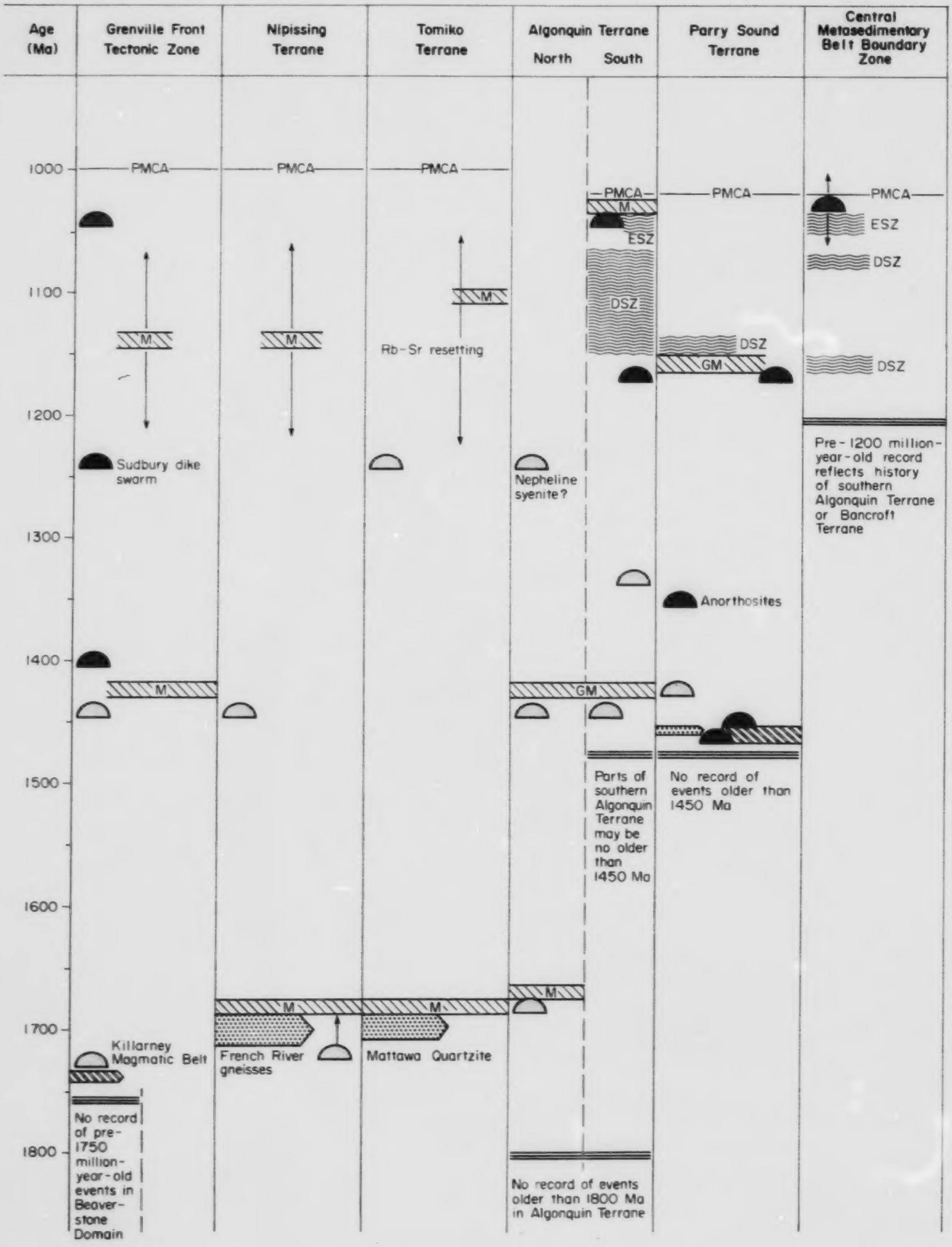

Age (Ma)
Grenville Front Tectonic Zone
Nipissing Terrane
Tomiko Terrane
Algonquin Terrane North South
Parry Sound Terrane
Central Metasedimentary Belt Boundary Zone
1000
1100
1200
1300
1400
1500
1600
1700
1800
PMCA
M
GM
ESZ
DSZ
Rb-Sr resetting
Sudbury dike swarm
Nepheline syenite?
Anorthosites
Pre-1200 million-year-old record reflects history of southern Algonquin Terrane or Bancroft Terrane
Parts of southern Algonquin Terrane may be no older than 1450 Ma
No record of events older than 1450 Ma
Killarney Magmatic Belt
French River gneisses
Mattawa Quartzite
No record of pre-1750 million-year-old events in Beaverstone Domain
No record of events older than 1800 Ma in Algonquin Terrane

**Figure 25.17.** Sketch summarizing time-space relationships within the Central Gneiss Belt.

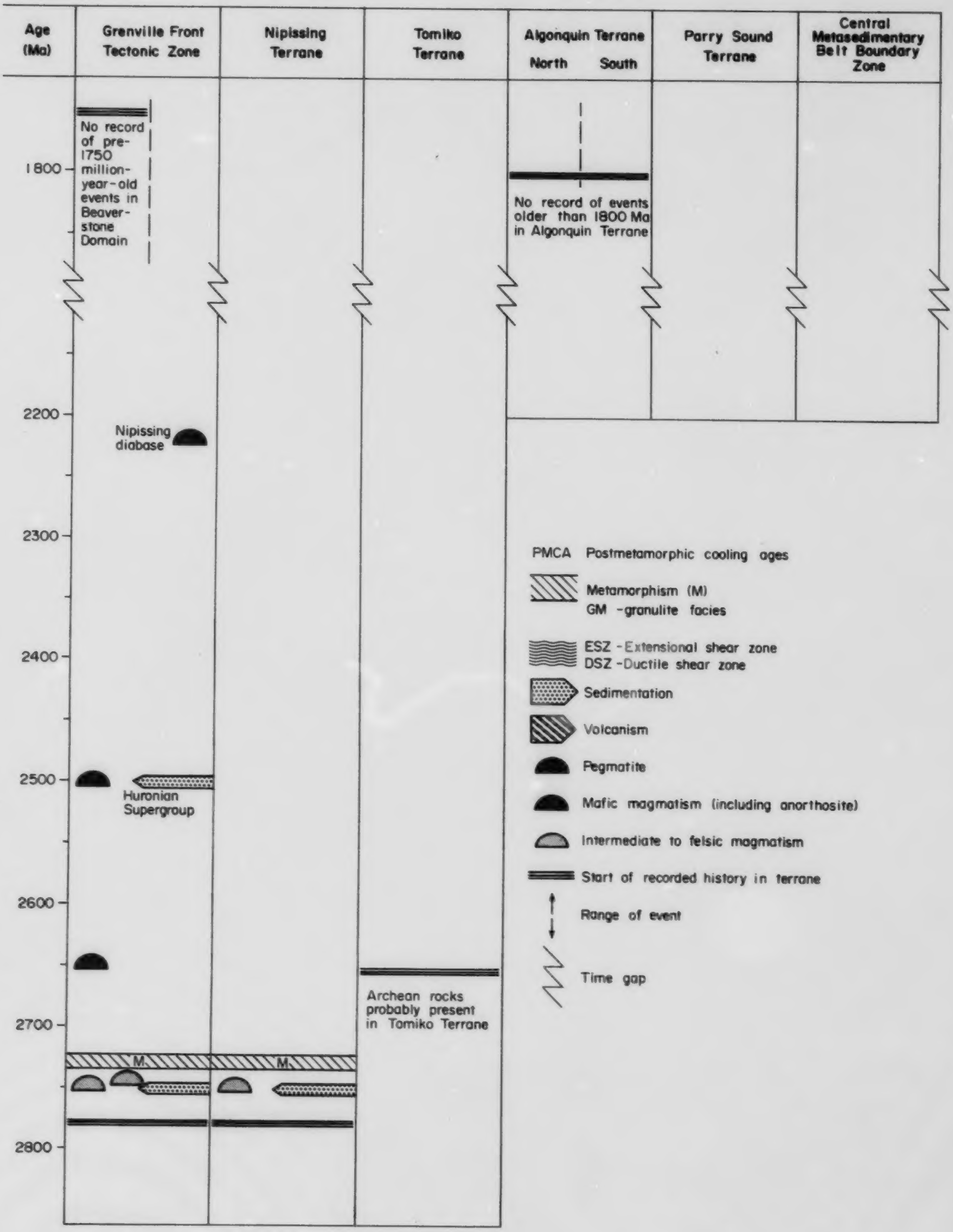


**Figure 25.17.** Sketch summarizing time-space relationships within the Central Gneiss Belt.

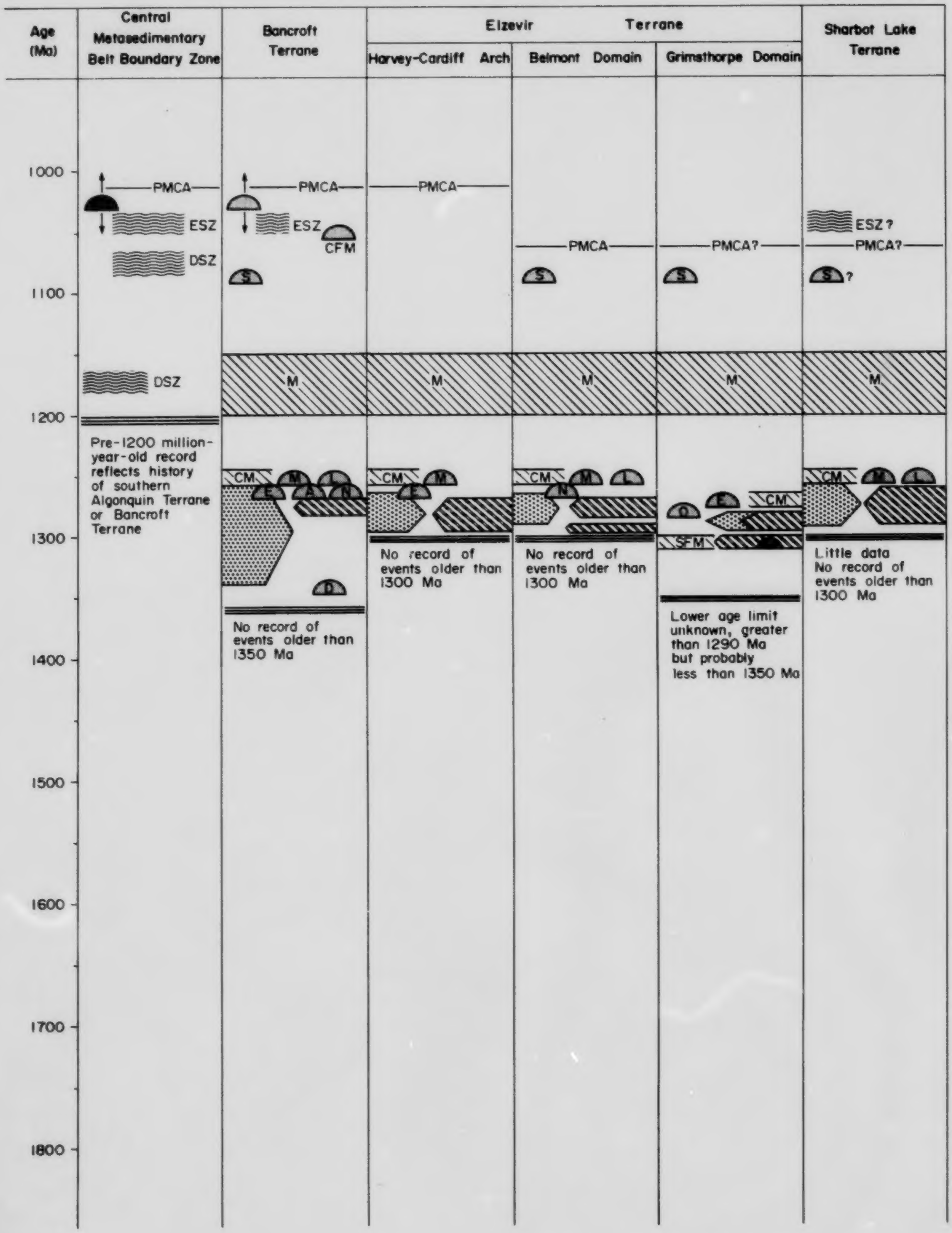

Age (Ma)
Central Metasedimentary Belt Boundary Zone
Bancroft Terrane
Elzevir Terrane
Harvey-Cardiff Arch
Belmont Domain
Grimsthorpe Domain
Sharbot Lake Terrane
1000
1100
1200
1300
1400
1500
1600
1700
1800
PMCA
ESZ
DSZ
DSZ
Pre-1200 million-year-old record reflects history of southern Algonquin Terrane or Bancroft Terrane
PMCA
ESZ
CFM
S
M
CM
M
L
E
A
N
D
No record of events older than 1350 Ma
PMCA
M
CM
M
E
No record of events older than 1300 Ma
PMCA
S
M
CM
M
L
N
No record of events older than 1300 Ma
PMCA?
S
M
D
E
CM
SFM
Lower age limit unknown, greater than 1290 Ma but probably less than 1350 Ma
ESZ ?
PMCA?
S ?
M
CM
M
L
Little data
No record of events older than 1300 Ma

**Figure 25.18.** Sketch summarizing time-space relationships within the Central Metasedimentary Belt.

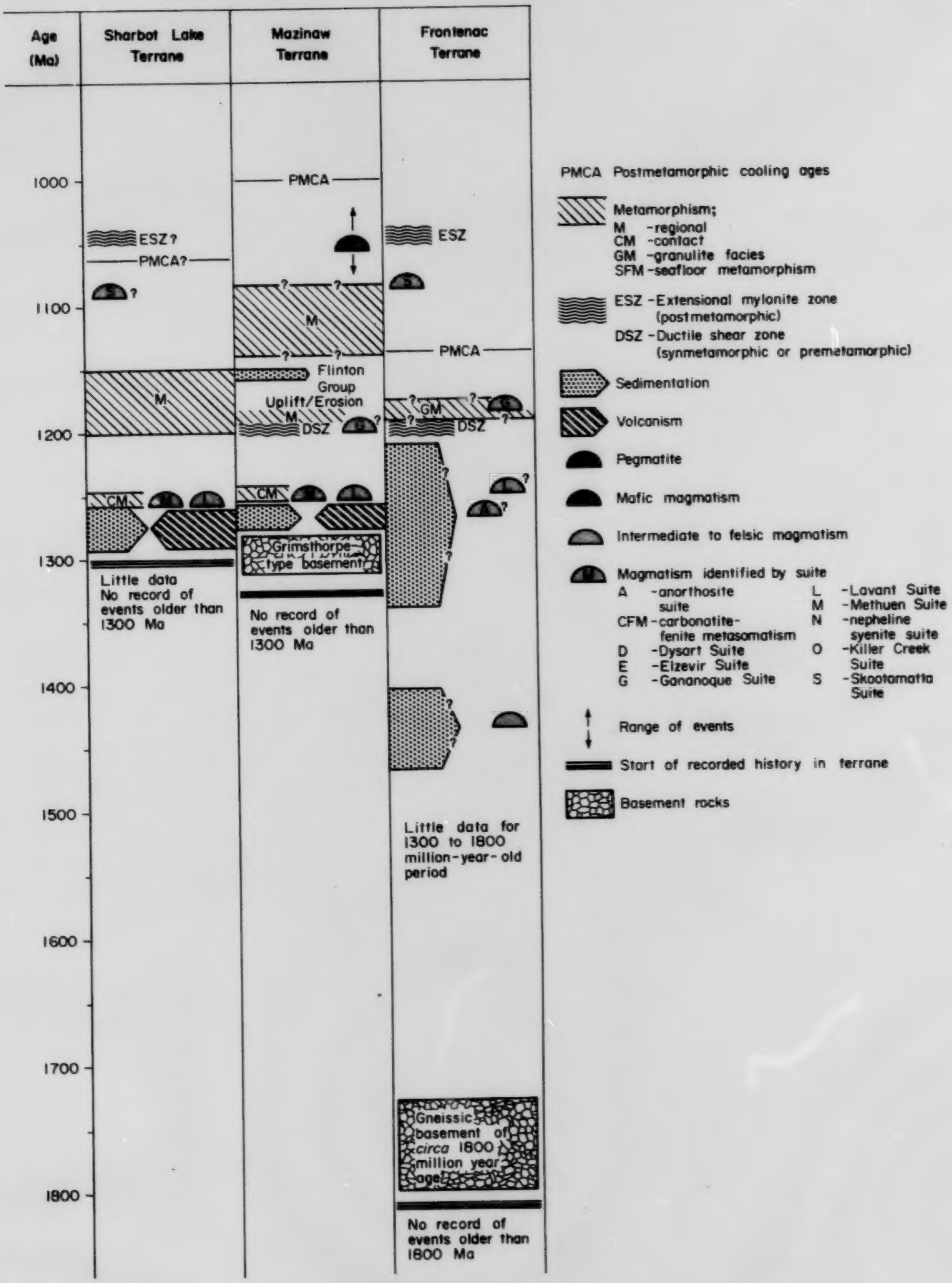

Age (Ma)
Sharbot Lake Terrane
Mazinaw Terrane
Frontenac Terrane
1000
1100
1200
1300
1400
1500
1600
1700
1800
PMCA
ESZ?
PMCA?
S ?
M
CM
Little data No record of events older than 1300 Ma
?
Flinton Group
Uplift/Erosion
DSZ
Grimsthorpe-type basement
No record of events older than 1300 Ma
ESZ
GM
Little data for 1300 to 1800 million-year-old period
Gneissic basement of circa 1800 million year age
No record of events older than 1800 Ma
PMCA Postmetamorphic cooling ages
Metamorphism;
M -regional
CM -contact
GM -granulite facies
SFM -seafloor metamorphism
ESZ -Extensional mylonite zone (postmetamorphic)
DSZ -Ductile shear zone (synmetamorphic or premetamorphic)
Sedimentation
Volcanism
Pegmatite
Mafic magmatism
Intermediate to felsic magmatism
Magmatism identified by suite
A -anorthosite suite
CFM -carbonatite-fenite metasomatism
D -Dysart Suite
E -Elzevir Suite
G -Gananoque Suite
L -Lavant Suite
M -Methuen Suite
N -nepheline syenite suite
O -Killer Creek Suite
S -Skootamatta Suite
Range of events
Start of recorded history in terrane
Basement rocks

**Figure 25.18.** Sketch summarizing time-space relationships within the Central Metasedimentary Belt.

ism (Lumbers et al. 1990). These domains and assemblages may represent a relatively intact sequence with only minor thrusting exposing different structural levels, or a sequence of accreted terranes which may have been initially separated from one another by hundreds of kilometres (perhaps as different sections of a more extensive continental margin, e.g., analogous to the development of the Cordillera during the Neoproterozoic to Ordovician). The amalgamation of the Bancroft and Elzevir terranes into a coherent "superterrane" was largely complete by ca. 1250 to 1240 Ma, as the Bancroft, Elzevir and, in part, the Sharbot Lake Terrane contain the same plutonic suites (Nepheline Syenite, Killer Creek, Anorthosite, Elzevir, Lavant, Methuen), with the Lavant and Methuen suites representing the crustal stabilization event. Although these terranes may have been accreted to one another by this time, this resulting "superterrane" may still have been separate from Laurentia. The setting of the Bancroft and Elzevir terranes in the 1300 to 1240 million-year-old interval may be analogous to the present setting of the Sunda and Banda arcs (*see* Easton, this volume).

Mazinaw Terrane history is less clear. The terrane consists of a Grenville Supergroup volcanic arc (ca. 1250 Ma?), and a related siliciclastic and carbonate sediment sequence, probably constructed on Grimsthorpian-type basement rocks (ca. 1270 million-year-old Elzevir Suite intrusions). It was then intruded by Methuen Suite intrusions, such as the 1241 million-year-old Addington Granite. At that point, the history of the terrane diverged from that of the Elzevir and Bancroft terranes, for it was subsequently subjected to a high-grade (possibly granulite) metamorphic event, rapidly uplifted and eroded, and overlain by the Flinton Group sediments. These in turn were metamorphosed by a regional metamorphic event, dominated by high temperatures and slow cooling with minimal

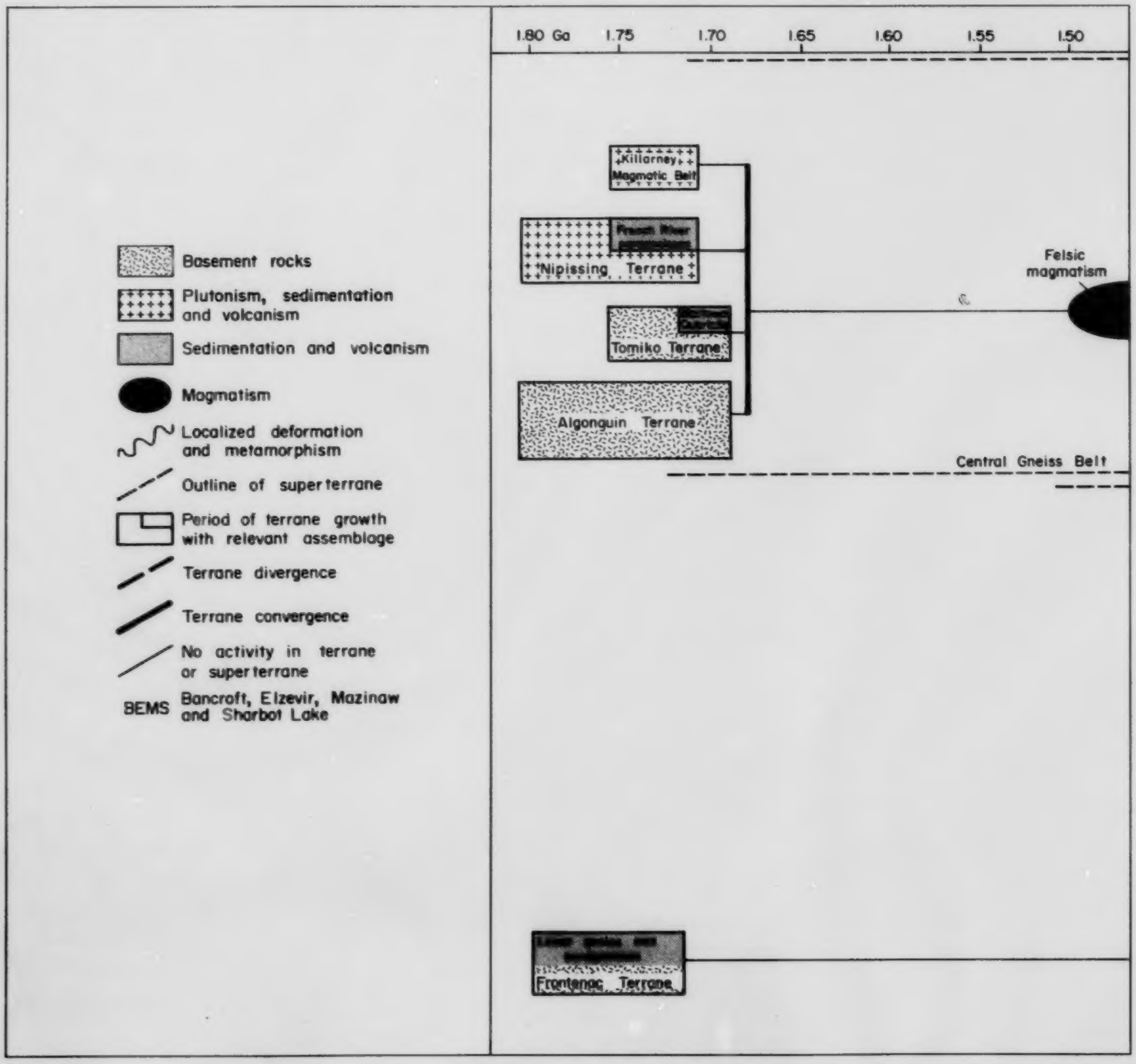

1.80 Ga
1.75
1.70
1.65
1.60
1.55
1.50
Killarney Magmatic Belt
French River
Nipissing Terrane
Tomiko Terrane
Algonquin Terrane
Felsic magmatism
Central Gneiss Belt
Frontenac Terrane
Basement rocks
Plutonism, sedimentation and volcanism
Sedimentation and volcanism
Magmatism
Localized deformation and metamorphism
Outline of superterrane
Period of terrane growth with relevant assemblage
Terrane divergence
Terrane convergence
No activity in terrane or superterrane
BEMS
Bancroft, Elzevir, Mazinaw and Sharbot Lake

**Figure 25.19.** Cladogram illustrating the tectonic evolution of the Grenville Province in Ontario.

compression. The early supracrustal and plutonic rocks are roughly analogous to the Elzevir Terrane, but the metamorphic and deformation history and style are similar to that of the Frontenac Terrane. The Mazinaw Terrane includes a calc-alkalic volcanic sequence, in contrast to the predominantly tholeiitic sequences within the Elzevir Terrane. Many granitoid intrusions in the Mazinaw Terrane are undated, and 1180 to 1160 million-year-old plutons of the Gananoque Suite may be present within the Mazinaw Terrane.

Carbonate and clastic sedimentation during the interval 1300 to 1240 Ma may have occurred in the Frontenac Terrane (Upper Marble and Quartzite unit).

The Central Gneiss Belt also locally exhibits evidence of magmatic activity in the 1300 to 1240 million year interval, possibly in an extensional regime. Events include emplacement of the 1238 million-year-old (Krogh et al. 1987) Sudbury Diabase dike swarm, emplacement of the Mulock Granite in the Tomiko Terrane and possible nepheline syenite suite magmatism of this age in the Nipissing and Algonquin terranes. These first 2 well-dated events correspond in age to Lavant and Methuen suite magmatism, which occurred in a postvolcanic, extensional regime. The nepheline syenite magmatism could be of the same age as that within the Central Metasedimentary Belt and, thus, would be slightly older (ca. 1290 to 1250 Ma). Note, the presence of similarly aged magmatic events does not necessarily imply that the Central Metasedimentary Belt and Central Gneiss Belt were contiguous at 1240 Ma.

## 1180 to 1160 Million-Year-Old Activity

Not until *circa* 1180 to 1160 Ma are congruent events recorded throughout the Grenville Province in Ontario.

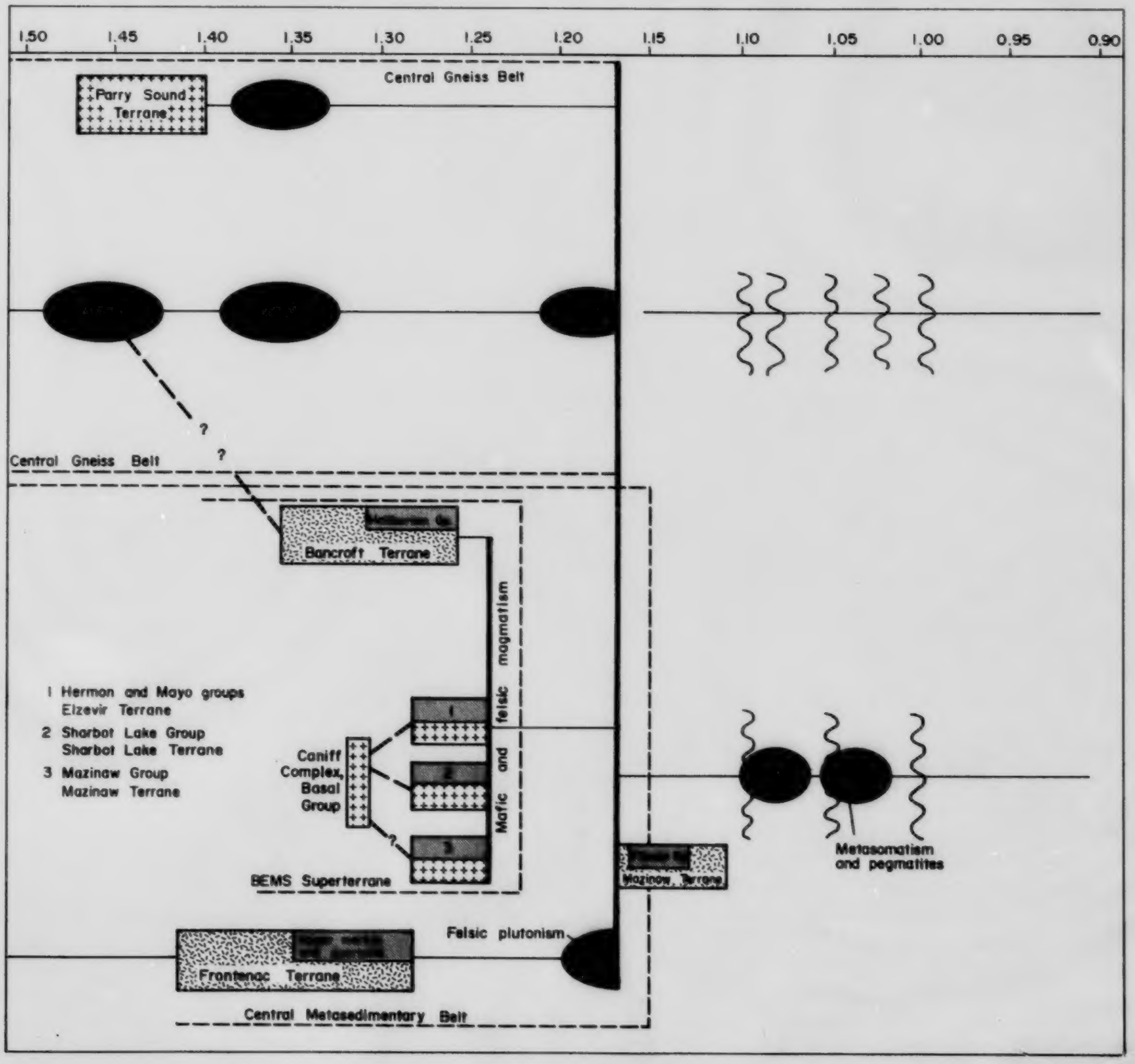

1.50
1.45
1.40
1.35
1.30
1.25
1.20
1.15
1.10
1.05
1.00
0.95
0.90
Central Gneiss Belt
Parry Sound Terrane
?
?
Central Gneiss Belt
Bancroft Terrane
Mafic and felsic magmatism
1 Hermon and Mayo groups
Elzevir Terrane
2 Sharbot Lake Group
Sharbot Lake Terrane
3 Mazinaw Group
Mazinaw Terrane
Caniff Complex, Basal Group
1
2
3
?
BEMS Superterrane
Mazinaw Terrane
Metasomatism and pegmatites
Felsic plutonism
Frontenac Terrane
Central Metasedimentary Belt

**Figure 25.19.** Cladogram illustrating the tectonic evolution of the Grenville Province in Ontario.

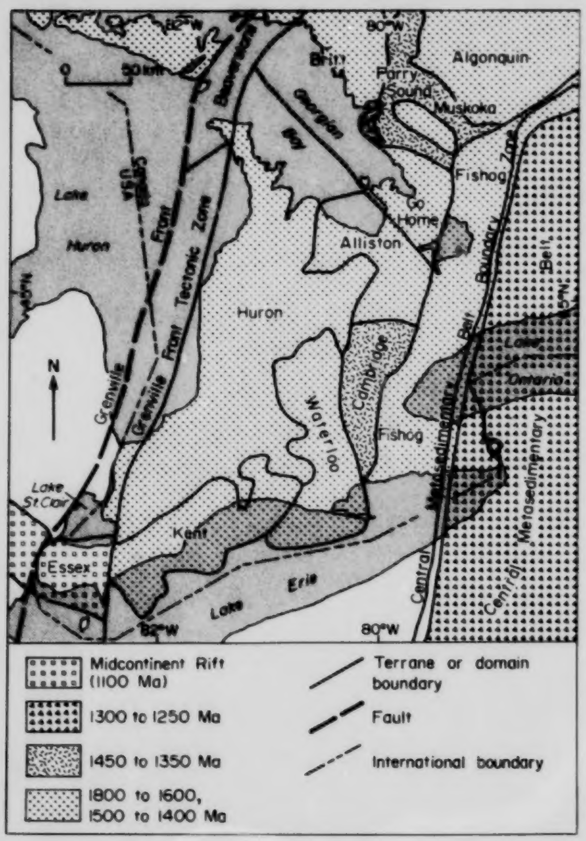


**Figure 25.20.** Lithotectonic domains in the Precambrian basement of southwestern Ontario interpreted from drill core and cuttings and published aeromagnetic maps. Patterns refer to inferred crustal ages. *Modified from* Carter and Easton (1990).

Thrusting of the Parry Sound Terrane over the Algonquin Terrane and granulite-facies metamorphism in the Parry Sound Terrane occurred at *circa* 1160 Ma (van Breemen et al. 1986). Earliest recorded deformation in the Central Metasedimentary Belt Boundary Zone occurred at *circa* 1180 Ma (McEachern 1990). Within the Central Metasedimentary Belt, Lumbers et al. (1990) suggest that the main period of regional metamorphism related to the "Grenville Orogeny" occurred between 1240 and 1180 Ma based on ages of plutonic suites in the Central Metasedimentary Belt. This is consistent with observations in the Mazinaw Terrane that high-grade metamorphism occurred prior to deposition of the *circa* 1155 million-year-old Flinton Group, and that this earlier metamorphic event overprinted pre-Flinton thrusts in the terrane (Easton and Ford 1991). Within the Elzevir Terrane, a thermal pulse at *circa* 1140 Ma (e.g., York 1984) interrupted the cooling history of the Elzevir Terrane after the emplacement of Lavant Suite diorites and gabbros. Within the Frontenac Terrane, Gananoque Suite magmatism is coincident with granulite-facies metamorphism between 1175 and 1160 Ma (Marcantonio et al. 1990; van Breemen and Davidson 1988b; Lumbers et al. 1990). All of these ages are consistent with a major orogenic period (referred to as $M_1$ and $D_1$) extending from Parry Sound to Kingston at 1180 Ma. This event corresponds roughly to the Elzevirian Orogeny (Baer 1981; Moore and Thompson 1980; Easton 1986). The cause of this regional deformation and metamorphism is not known. In Figure 25.19, it is assumed that it was caused by collision between the Central Gneiss Belt and a continental block with a platform consisting of the Frontenac and possibly the Central Granulite terranes. The range in age for this orogenic activity, as well as the scale of crustal thickening and heating of the crust through magmatic activity to generate widespread granulite-facies conditions across the Central Gneiss Belt and the Central Metasedimentary Belt, suggests a continental collisional event, rather than an arc-arc collision.

The Mazinaw and Frontenac terranes may have become accreted to one another during this orogenic episode, as both seem to record similar metamorphic and deformational histories related to the Elzevirian Orogeny. The history of the 2 terranes during the Ottawan Orogeny is not the same, however, as the Mazinaw Terrane was subjected to an amphibolite-facies metamorphic event not preserved in the Elzevir Terrane. This difference might be attributed to thicker crustal stacking over the Mazinaw Terrane during the Ottawan Orogeny, resulting in deeper burial.

Knowledge of events within the Central Metasedimentary Belt between *circa* 1160 and 1090 Ma is limited. In the Mazinaw Terrane, rapid uplift and erosion occurred, followed by Flinton Group deposition in a probable extensional environment. Little other activity is known from either the Central Gneiss Belt or the Central Metasedimentary Belt. Perhaps it was a period of uplift and stabilization.

## 1100 to 1060 Million-Year-Old Activity

Between *circa* 1100 and 1000 Ma, 2 discrete tectonic events appear to closely follow one another. The first, between 1100 and 1060 Ma, resulted in thrusting within the Central Gneiss Belt and along the Central Metasedimentary Belt Boundary Zone that is recorded by syntectonic pegmatites ($D_2$)(e.g., Nadeau and van Breemen 1990; McEachern 1990; van Breemen and Hanmer 1986). This thrusting event corresponds roughly to the "Ottawan" Orogeny. Granulite-facies metamorphism occurred in the central Algonquin Terrane (van Breemen et al. 1986), and retrograde (upper amphibolite) metamorphism is recorded in terrane and domain boundaries. Within the Mazinaw Terrane, upper greenschist- to lower amphibolite-facies metamorphic conditions existed, as the Flinton Group was metamorphosed in the interval *circa* 1140 to 1050 Ma. The regional metamorphism ($M_2$) in the Mazinaw Terrane has no known associated magmatism and, thus, was probably induced by burial, probably caused by crustal stacking during the same thrusting event that affected the Central Gneiss Belt. The cause of this event is unknown. It could be due to continent-continent collision south of the exposed Grenville Province. Other than in the Algonquin and Mazinaw terranes, there is no indication of a significant regional metamorphic event in the Grenville Province during this period, although a low grade thermal event at this time may have reset Rb-Sr and K-Ar isotopic systems across the Grenville Province (e.g., ubiquitous 1070 million-year-old titanite ages in the Belmont Domain, Davis and Bartlett 1988).

The second event is represented in the Central Metasedimentary Belt by *circa* 1090 to 1075 million-year-old Skootamatta Suite postmetamorphic intrusions emplaced into the Bancroft, Elzevir and Frontenac terranes (Corriveau et al. 1990). Postmetamorphic Skootamatta Suite intrusions have ages of *circa* 1088 Ma, implying that the *circa* 1070 million-year-old titanite ages in the Belmont Domain record cooling, not another thermal pulse. Corriveau et al. (1990) suggest that the ultrapotassic to shoshonitic Skootamatta Suite plutons were emplaced above an active subduction zone that dipped to the northwest beneath the Elzevir Terrane. Magmatism related to this arc, however, consists only of a limited number of plutons, and no other major magmatic event is known from either the Central Metasedimentary Belt or the Central Gneiss Belt in this interval. Shoshonitic to ultrapotassic magmatism is known from recent posttectonic extensional environments, and in these instances, the geochemical characteristics of these intrusions were inherited from mantle subjected to an earlier subduction-related event (in this case, ca. 1270 to 1245 Ma; Corriveau et al. 1990). This latter explanation is preferred.

Midcontinent Rift magmatism occurred in the foreland region of the Grenville Orogen at this time (1.1 to 1.08 Ga), although it cannot be related to the ultrapotassic magmatism in the Central Metasedimentary Belt. Anderson and Burke (1983) and Gordon and Hempton (1986) have postulated that the Midcontinent Rift represents an "impactogen" that formed as a result of continental collision, analogous to the Red Sea developing in front of the Zagros Front in the Middle East (*see* Easton, this volume).

It was during this interval that the architecture of the Central Metasedimentary and Central Gneiss belts as observed today was developed. Ductile thrusting resulted in stacking and disruption of early formed "superterranes." The Elzevir, Bancroft and Sharbot Lake terranes were thrust over the Frontenac and Mazinaw terranes. Subsequent extension along fault boundaries resulted in considerable vertical movement and local rapid erosion, which resulted in the terrane and domain boundaries currently observed in the Grenville Province. For example, the Frontenac Terrane was probably unroofed and uplifted rapidly, whereas the Mazinaw Terrane was buried longer, and uplifted slowly, thereby suffering more intense metamorphism during the Ottawan Orogeny.

## 1060 to 1000 Million-Year-Old Activity

Following the compressional activity between 1100 and 1060 Ma, extensional events occurred throughout the Central Metasedimentary Belt for roughly 60 million years, particularly in the Bancroft Terrane, manifested by carbonatite- fenite suite magmatism and metasomatism and granite pegmatite injection (Lumbers et al. 1990; Lentz 1991). During this interval, the area cooled at different rates across the Central Metasedimentary Belt and the Central Gneiss Belt, the difference in cooling rates resulting from extensional gravitational collapse of the thrust stack assembled during the Elzevirian and Ottawan orogenies (cf., Cosca 1989; Mezger et al. 1991). By *circa* 1000 Ma, cooling below hornblende argon-blocking temperatures was complete across most of the Grenville Province, with cooling through biotite blocking temperatures complete by 950 Ma. The timing of the "Grenville Event" or "Grenville Orogeny" has long been associated with this late, cooling event, which in essence reflects the end of regional orogenic activity in the Grenville Province, not the peak of orogenesis.

Extensional faulting along domain and terrane boundaries in the Central Gneiss Belt occurred in this period (ca. 1040 Ma, Nadeau and van Breemen 1990), and major shear systems such as the Robertson Lake mylonite zone and the Bancroft shear zone developed. The Sharbot Lake Terrane was probably emplaced at this time as a thin crustal sheet (cf., Real and Thomas 1987), representing a sliver of Elzevir Terrane, that was back thrust and wedged between the Mazinaw and Frontenac terranes.

## Post–1000 Million-Year-Old Activity

Extension continued in the period 1000 to 900 Ma during cooling of the Grenville Orogen, and at *circa* 900 Ma was characterized by mafic dike injection (Kingston Swarm), minor carbonatitic and alkalic magmatism, and faulting related to the development of a precursor to the Ottawa graben.

From 1000 to 500 Ma, the Grenville Orogen was uplifted and eroded. Much of the detritus from the orogen may have been transported northward into Arctic region sedimentary basins, such as the Neoproterozoic Shaler Basin (e.g., Rainbird et al. 1991). By the Early Paleozoic, the Grenville Orogen was essentially a peneplain.

Extension was renewed in the Neoproterozoic to Early Cambrian, with faulting along the Ottawa–Bonnechere Graben coincident with the emplacement of carbonatitic and alkalic complexes in the Nipissing area and Grenville Swarm diabase dikes. This extensional event is probably related to rifting of North America in conjunction with the development of the Iapetus Ocean. During this time, and until the deposition of fluvial and nearshore marine sediments in the Middle Cambrian, the area was in a tropical environment; and regoliths and karsts were developed on the exposed Grenville surface.

## MINERAL DEPOSITS

Many of Ontario's early mines were located in the Central Metasedimentary Belt, and significant production of iron, zinc, molybdenum, uranium and lead occurred, primarily at the turn of the century. Copper, nickel and gold prospects are also common in the Central Metasedimentary Belt, and limited gold production has occurred (Carter 1984). In addition, a variety of industrial minerals have been, and are being, produced from the Grenville Province, primarily within the Central Metasedimentary Belt, including graphite, nepheline syenite, talc, tremolite, calcite and dolomite marbles, roofing granules, flagstone and building stone.

The mineral potential of the Central Gneiss Belt is difficult to evaluate. Paragneiss-rich domains within the Algonquin Terrane host graphite deposits and potentially host sediment-associated, exhalative (SEDEX) base-metal deposits. The Parry Sound Domain, with its abundance of mafic metaigneous rocks and paragneisses, hosts numerous gold and copper showings. Industrial mineral potential of the Central Gneiss Belt includes flagstone and building stone resources, graphite and the use of anorthosite as an alumina source, as outlined by Easton and Fyon (this volume).

The mineral potential of the Central Metasedimentary Belt is significantly higher, although, as outlined in Easton and Fyon (this volume), some commodities are concentrated in specific terranes or domains, a reflection of the stratigraphy and tectonic diversity of the region. Mineralization within the Central Metasedimentary Belt appears to have occurred primarily in extensional regimes (*see* Easton and Fyon, this volume) during 4 periods, as follows: 1) copper-nickel-platinum group element mineralization in *circa* 1290 to 1270 million-year-old mafic intrusions, probably formed during oceanic crust generation in a marginal basin environment; 2) iron-titanium, iron-skarn, iron-sulphide, iron formation, stratiform carbonate-hosted zinc and minor base metal mineralization related to volcanism, sedimentation and emplacement of Lavant and Methuen suite gabbroic and granitic plutons in the *circa* 1255 to 1240 million-year-old interval, again, probably in a marginal basin. Many important industrial mineral deposits developed at this time in the contact aureoles of Lavant and Methuen suite plutons; 3) gold mineralization, possibly related to local extensional environments that developed during and after the Elzevirian Orogeny at *circa* 1180 to 1155 Ma. Graphite deposits in the Central Metasedimentary Belt may also have formed at this time; and 4) uranium, rare-element and molybdenum deposits related to granite pegmatite injection and regional metasomatism in the Bancroft Terrane after the Ottawan Orogeny at *circa* 1060 to 1030 Ma. In addition to these 4 main mineralizing periods, a variety of industrial mineral deposits in the Mazinaw Terrane formed during Ottawan Orogeny regional metamorphism at *circa* 1150 to 1090 Ma.

To date, most mineral deposits, including past and present producers, found within the Grenville Province are small in comparison to other regions (e.g., Superior Province). We known of no geologic reason for the absence of large ore bodies from the Grenville Province, and it is our view that significant deposits remain to be discovered.

# PART 4: PALEOZOIC AND MESOZOIC SEDIMENTATION: TECTONIC INFLUENCES ON A "STABLE" CRATON

**D.K. Armstrong**

## INTRODUCTION

In this section, the tectonic evolution of the Paleozoic and Mesozoic eras in Ontario is reviewed. The interior of the North American craton, where Ontario is situated, has traditionally been considered to have been a stable platform during most of the Phanerozoic. However, the sedimentary record preserved on the craton provides abundant evidence of crustal tectonism. Subsidence histories of intracratonic basins (e.g., Sloss 1988a) and the related patterns of sedimentation and erosion (*see* Johnson et al., this volume), local structural controls on sedimentation and diagenesis (e.g., Sanford et al. 1985), and recent intraplate seismicity (e.g., Fakundiny et al. 1978) indicate that the craton was (and still is to some extent) a tectonically active element.

This section of Chapter 25 reviews the tectonic regime of the eastern part of the North American craton during the Paleozoic and Mesozoic eras, providing a large-scale context for discussion of the relationships between tectonics and the sedimentary record as described by Johnson et al. (this volume). The implications for models of sedimentary basin evolution and for our understanding of the distribution of mineral resources in these basins are also discussed.

## EVOLVING TECTONIC REGIME OF EASTERN NORTH AMERICA

Tectonically, Ontario is situated in an intraplate setting with its closest and most influential continental plate margin located along the eastern (present orientation) margin of North America (Bally et al. 1989). The tectonic regime along this plate margin during the Phanerozoic can be described in terms of 2 tectonic cycles, termed "Wilson cycles" (*see* Hatcher 1989; Sanford et al. 1985). The first tectonic cycle commenced with rifting during the latest part of the Neoproterozoic to the Early Cambrian, and continued with the development of a passive continental margin open to the newly formed Iapetus Ocean. An active phase of collisional tectonism began during the Ordovician and continued until the late Paleozoic, when the African plate collided with North America, closing the Iapetus Ocean (Bally et al. 1989). The second tectonic cycle began with rifting during the latest Paleozoic and earliest Mesozoic (Hatcher 1989), which created the Atlantic Ocean and the passive continental margin which exists today.

During the active phase of the first tectonic cycle (i.e., throughout most of the Paleozoic), collision of various terranes, and ultimately the African plate, with the eastern margin of the North American plate produced the Appalachian Orogen. Collision of these terranes produced 3 major orogenies (Hatcher 1989): 1) the Middle(?) Ordovician to Late(?) Silurian Taconic Orogeny; 2) the Middle(?) Devonian to Early Mississippian(?) Acadian Orogeny; and 3) the Late Mississippian(?) to Permian Alleghanian Orogeny. Other possible minor orogenic events occurring along this orogenic belt include the Early Ordovician Penobscottian Orogeny and the Late Silurian Salinic Disturbance (Hatcher

1989). The uncertainty expressed in the age ranges assigned to each orogeny is a function of dating techniques (e.g., dates of orogenic intrusions, metamorphic events, unconformities associated with uplift, and clastic wedges shed from orogenic highlands), the diachronous nature of collisional events along the orogen (Hatcher 1989), and the inclusion, by some workers, of minor or local orogenic events as early or late phases of major events (e.g., Hatcher (1989) included the Salinic Disturbance as part of the Taconic Orogeny).

During the Mesozoic, Ontario was within an overall extensional regime related to the opening of the Atlantic Ocean. Extensional tectonism was manifested in Ontario by faulting, graben development and extension-related alkalic and kimberlitic plutonism in the Lake Timiskaming and Ottawa–Bonnechere grabens (Brummer 1978; *see* Sage, this volume) (Figure 25.21). However, Paleozoic age alkalic-carbonatite intrusions in the Ottawa–Bonnechere Graben (*see* Sage, this volume) indicate that it must have been a pre–Mesozoic structure. Mesozoic intrusions are also known from the Moose River Basin area and southeastern Ontario. More significant igneous intrusions related to Mesozoic rifting occur in Quebec and New England (McHone and Butler 1984).

Subsequent to rifting, divergent tectonism (i.e., mid-ocean ridge spreading) created the Atlantic Ocean and the stress regime in the North American plate became compressional. This compressional regime, with maximum principal horizontal stress oriented northeast to east-northeast (Zoback et al. 1986), is evidenced by significant, historical, intraplate seismic activity (Fakundiny et al. 1978; Bally et al. 1989).

## TECTONIC ELEMENTS IN ONTARIO

Deposition during the Paleozoic and Mesozoic occurred in 4 sedimentary basins in, and immediately adjacent to, Ontario (*see* Figure 25.21), on interbasinal shelf or arch areas and at times, as indicated by erosional outliers on the Precambrian Shield, across most of the continent. A structure contour map of the Precambrian basement surface (*see* Figure 1, Map 2582, map case) illustrates the sedimentary basins and intervening arches forming the tectonic elements of the province. The significant tectonic elements in and adjacent to Ontario are:

1. Appalachian Orogen: an orogenic belt formed during the Paleozoic era, along the eastern (present orientation) margin of the North American continental plate, by collisional tectonism (i.e., orogenic tectonism).
2. Appalachian Basin: an elongate foreland basin, developed on the western flank of the Appalachian Orogen, in response to lithospheric loading by thrust sheets emplaced during collisional tectonism (Quinlan and Beaumont 1984). Dominant sediment infill is orogen-derived clastic sediments.
3. Michigan, Moose River and Hudson Bay basins: all are roughly circular, saucer-shaped intracratonic basins. These basins have considerably thinner sedimentary successions preserved, relative to the Appalachian Basin, and their sediment infill is dominated by carbonates, evaporites and craton-derived clastic sediments. The origin and development of intracratonic basins is discussed below.
4. Arches: basement-cored, structural highs, separating sedimentary basins. A conjugate network of arches on a continental scale has been proposed by Sanford et al. (1985). Periodic uplift centred on these arches (i.e., cratonic tectonism or epeirogeny) during the Phanerozoic controlled sedimentation patterns, caused erosion of earlier sediments and of the basement, and formed the focus of unconformity development. The Cape Henrietta Maria Arch separates the Moose River Basin from the Hudson Bay Basin. The Algonquin and Findlay arches separate the Michigan and Appalachian basins.
5. Lake Timiskaming and Ottawa–Bonnechere grabens: the Lake Timiskaming graben probably developed in response to Mesozoic rifting and the opening of the Atlantic Ocean (Kumarapeli 1978), and preserves lower to middle Paleozoic sediments of the Lake Timiskaming outlier. The Ottawa–Bonnechere Graben preserves a number of lower Paleozoic outliers along the Ottawa River valley and Lake Nipissing areas. Mesozoic lamprophyric dikes are associated with both of these grabens; Paleozoic alkalic-carbonatite intrusions are associated with the Ottawa–Bonnechere Graben (*see* Sage, this volume).

## TECTONICS AND SEDIMENTATION

Johnson et al. (this volume) reviewed the Paleozoic and Mesozoic geology of Ontario and examined the tectonic controls on sedimentation by applying sequence stratigraphic concepts (Sloss 1963; Mitchum et al. 1977) to this sedimentary record. Johnson et al. (this volume) subdivided the Paleozoic and Mesozoic successions of the province into 10 depositional sequences which developed in response to a combination of tectonic forces and eustatic sea level fluctuations (Figure 25.22). The internal architecture of these sequences can largely be attributed to tectonic forces operating both at the craton margin (i.e., orogenic tectonism) and on the craton itself (i.e., cratonic tectonism).

The spatial and temporal distribution of these 10 depositional sequences (numbered 1 through 10) and their gross lithologic components are plotted on Chart D (*see* Map 2582, map case), which is summarized in Figure 25.22. The present and interpreted original extent of sedimentation for each sequence is plotted on the sequence distribution maps (*see* Figures 2 to 11, Map 2582, map case). Also plotted on these maps are the isopachous thicknesses of sediments which constitute each sequence and the lateral distribution of dominant sediment types within each sequence. The outcrop and subcrop (beneath Quaternary cover) distribution of the depositional sequences is shown on the Tectonic Assemblages of Ontario map (*see* Maps 2575 to 2578, map case, sequences labelled P1 through P10).

Orogenic tectonism is evidenced by 3 major pulses of orogen-derived clastic sediments in Sequences 3 and 4, 8, and 9, shed during the main phases the Taconic, Acadian and Alleghanian orogenies, respectively. The smaller clastic pulses of Sequences 2, 5 and 6 reflect minor orogenic activity which may represent discrete collisional events or early and/or late phases of the main orogenic events. There is no evidence of the Penobscottian Orogeny (Hatcher 1989) known in Ontario; however, sediments of this age (Late Cambrian to Early Ordovician—Sequence 1) are not well exposed in the province.

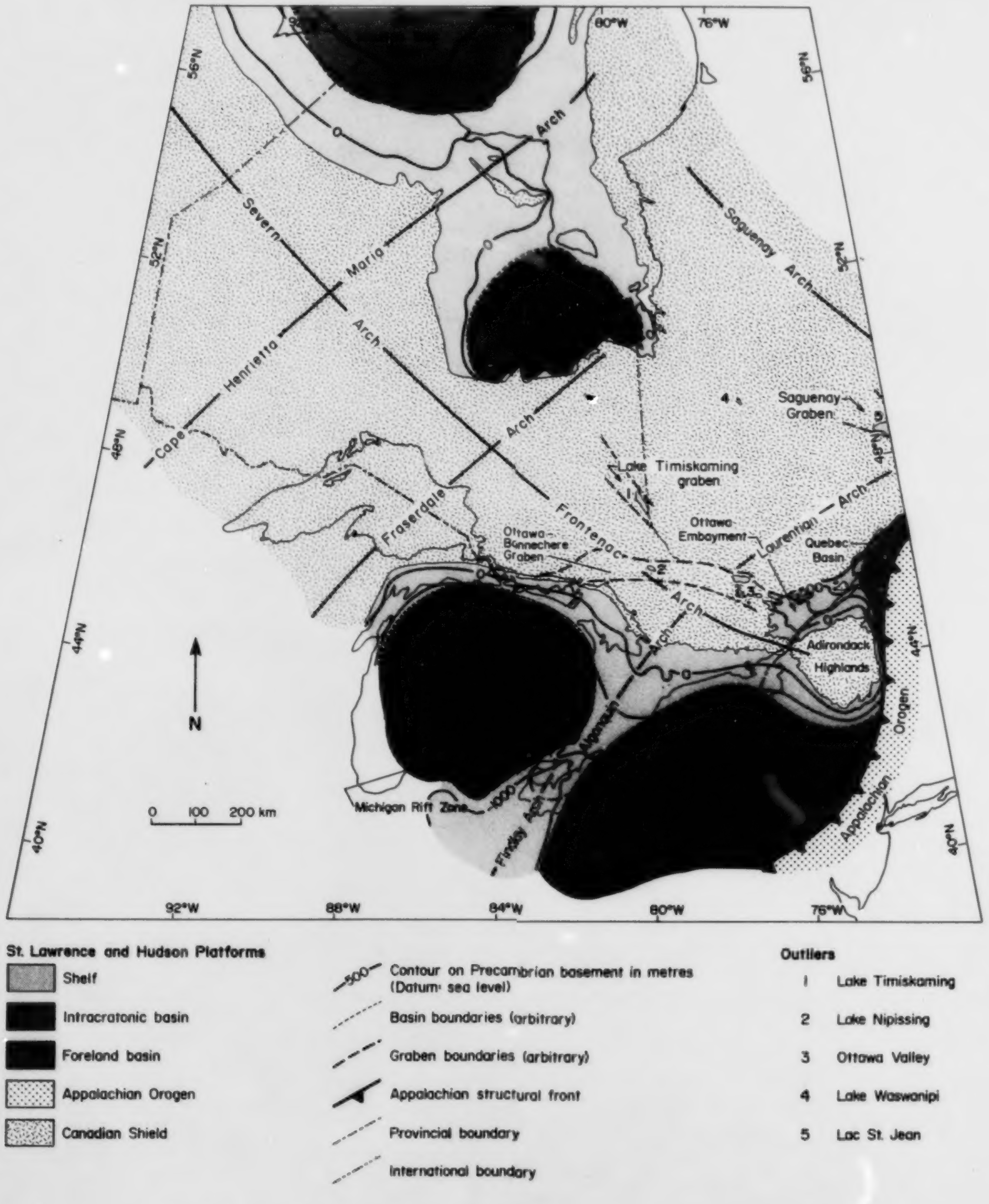

80°W
76°W
56°N
52°N
48°N
44°N
40°N
92°W
88°W
84°W
80°W
76°W
0
Cape Henrietta Maria Arch
Severn Arch
Saguenay Arch
Fraserdale Arch
Frontenac Arch
4
5
Saguenay Graben
Lake Timiskaming graben
1
Ottawa Embayment
Laurentian Arch
Quebec Basin
Ottawa-Bonnechere Graben
2
3
500
Adirondack Highlands
Algonquin Arch
Appalachian Orogen
Michigan Rift Zone
1000
Findlay Arch
N
0 100 200 km
St. Lawrence and Hudson Platforms
Shelf
Intracratonic basin
Foreland basin
Appalachian Orogen
Canadian Shield
500
Contour on Precambrian basement in metres (Datum: sea level)
Basin boundaries (arbitrary)
Graben boundaries (arbitrary)
Appalachian structural front
Provincial boundary
International boundary
Outliers
1 Lake Timiskaming
2 Lake Nipissing
3 Ottawa Valley
4 Lake Waswanipi
5 Lac St. Jean

**Figure 25.21.** Principal Paleozoic and Mesozoic tectonic elements of Ontario and adjacent regions.

Cratonic tectonism, such as intracratonic basin subsidence and arch-centred uplift, is evidenced in the shape, distribution and internal makeup of the 10 depositional sequences. Basin-centred subsidence is indicated in the isopach maps (*see* Figures 2 to 11, Map 2582, map case) of most sequences and by the basin-centred distribution of some sediments such as evaporites (e.g., Sequences 6 and 7) and certain carbonate facies (e.g., Sequence 5). The dominant type of clastic sediment in intracratonic basins is craton derived, generated by erosion of older sediments and basement material from uplifted adjacent arch areas. Uplift of the Algonquin Arch is indicated by the general restriction of orogen-derived clastic sediments to the Appalachian Basin and by the development of arch-centred unconformities. This last point is especially evident during the Silurian Sequences 4 and 5, as highlighted in the recent investigations of Brett et al. (1990).

The 10 depositional sequences identified by Johnson et al. (this volume), constituting the Paleozoic and Mesozoic succession in Ontario, are largely bounded by unconformities (*see* Figure 25.22; Chart D, Map 2582, map case). Unconformities develop in response to relative sea level drop, a function of eustatic sea level drop or tectonic uplift or a combination of both. Those sequence boundaries that correspond to Sloss's (1963) continental sequence boundaries (i.e., boundaries between Sequences 1 and 2, 6 and 7, and 9 and 10) have been variously attributed to eustatic sea level drops (Bally 1989) or to broad continental uplift (Sloss 1988a, 1991). The boundary between Sequences 3 and 4) coincides with a Late Ordovician glacio-eustatic sea level drop (Dennison 1976).

Note, however, that some Ontario sequence boundaries are not marked by unconformities (*see* Figure 25.22; Chart D, Map 2582, map case). There are 3 reasons for this. First, these sequences were selected to reflect significant changes in sediment type or depositional pattern. Thus, the boundary between Sequences 2 and 3 in southern Ontario reflects the influx of orogen-derived clastic sediments related to the main phase of the Taconic Orogeny and only locally is a disconformity recognized (*see* Johnson et al., this volume).

Second, some sequence-bounding unconformities may not be recognized because of the non-uniform exposure of units from basin to basin and throughout the stratigraphic column. For instance, more sequence boundaries may have been identified within the Cambrian interval (Sequence 1) if it were better exposed. Also, fewer sequence-bounding unconformities were identified in the northern Ontario basins. This again may be a function of the poor exposure of the entire succession in the north, and of our lack of knowledge concerning this area.

Third, some sequence-bounding unconformities may not exist in the Hudson Bay and Moose River basins because these unconformities were directly related to orogenic tectonism and these basins are farther from the source of that tectonism, that is, the Appalachian Orogen. Lithospheric and sedimentary responses to orogenic tectonism may be more subtle in these interior basins. This raises 2 additional points with potential significance to crustal evolution. First, the lack of interbasin correlation of some sequence boundaries may indicate which boundaries are tectonically induced and which are caused by eustatic sea level drop. Conversely, this lack of continuity may be indicating something about the evolution of the crust during the Paleozoic. Note on Chart D, Map 2582, that in the northern basins, more unconformities are recognized in the Devonian sequences (Sequences 7 and 8) than in the Ordovician and Silurian sequences (Sequences 2 through 6). This seems to support the suggestion (e.g., Klein 1991) that crustal rigidity increased with time and allowed the transmission of lithospheric flexure farther out onto the craton. Again however, the paucity and uncertainty of data from these northern basins precludes a definitive conclusion.

## MODELS OF INTRACRATONIC BASIN DEVELOPMENT AND CRATONIC TECTONISM

The bulk of the Paleozoic and Mesozoic sedimentary succession in Ontario was deposited in, or adjacent to, 3 intracratonic basins (the Michigan, Hudson Bay and Moose River basins), so an understanding of their evolution is critical to understanding the origin and distribution of resources within them. Understanding the geology of these basins is also critical to hydrogeological and environmental planning decisions involving these strata.

This section reviews the current models of intracratonic basin development, most of which are based on studies of the 4 main Paleozoic intracratonic basins on the North American craton, the Williston, Illinois, Michigan and Hudson Bay basins (Quinlan 1987; Bally 1989). Commonly, workers either include the Moose River Basin with the Hudson Bay Basin or ignore it altogether. Consequently, even less is known about the origin and development of this small and distinct basin.

The origin and evolution of intracratonic basins are poorly understood and the subject of considerable controversy (Quinlan 1987; Sloss 1988b; Bally 1989; Klein 1991). Comparisons of the 4 major intracratonic basins developed on the North American craton (Quinlan 1987; Bally 1989) yield some basic common characteristics. They are roughly circular, saucer-shaped, sedimentary basins, developed entirely on continental lithosphere, not adjacent to an orogenic belt. Typically their sediment infill consists of shallow-water carbonates, evaporites, and craton-derived clastic sediments. Most of these basins do not display extensional rifting structures, or if present, these features significantly predate intracratonic basin subsidence.

Data on intracratonic basins come primarily from hydrocarbon exploration drilling and geophysical surveys. These data are limited, especially with respect to subbasin crust characteristics and early Paleozoic sedimentary history, and typically, coverage is inconsistent from basin to basin or even within a basin. Based on these limited and variable data, numerous models have been developed to explain the formation of intracratonic basins (e.g., McKenzie 1978; Sloss and Speed 1974; Klein and Hsui 1987). Quinlan

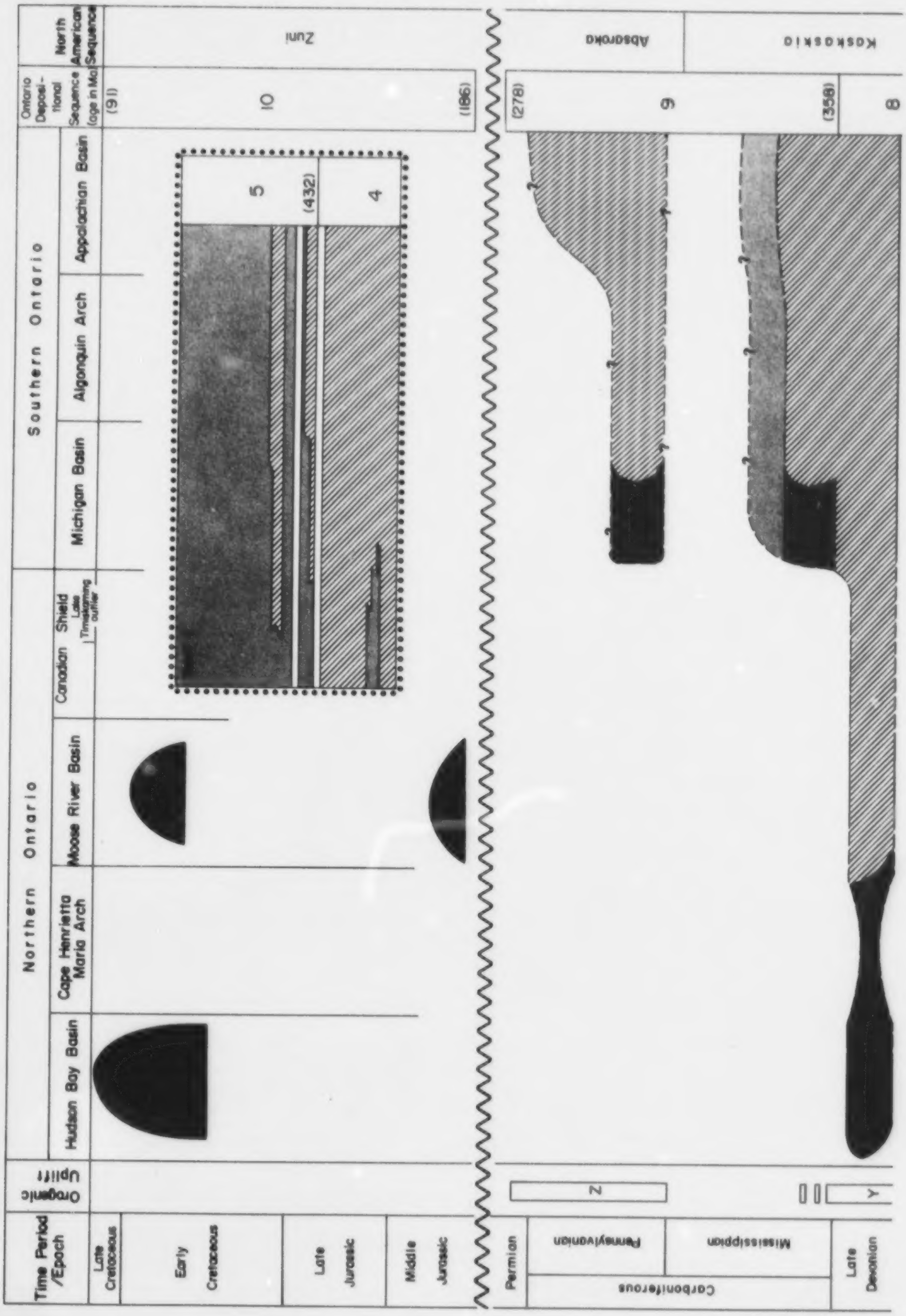


Figure 25.22. Condensed time-space diagram (*see* Chart D, Map 2582, map case) for Paleozoic and Mesozoic sequences of Ontario.

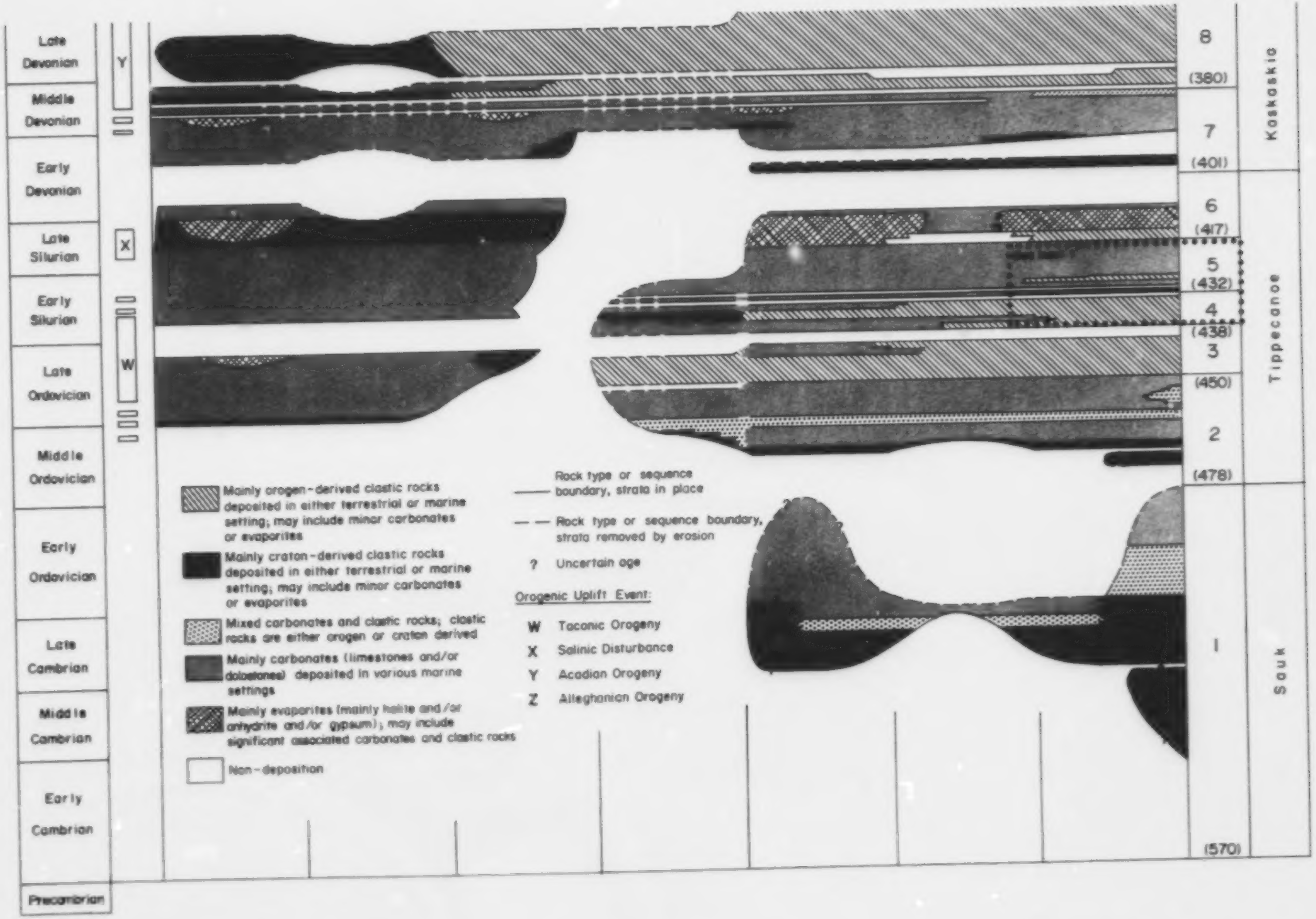


**Figure 25.22.** Condensed time-space diagram (*see* Chart D, Map 2582, map case) for Paleozoic and Mesozoic sequences of Ontario.

(1987) pointed out, however, that despite some similar characteristics, intracratonic basins exhibit significant differences. Others, such as Sloss (1988b) and Klein (1991), see significant commonality among these basins and interpret synchroneity in their development.

Synchroneity of basin initiation and subsidence amongst these basins may signify continental- or global-scale basin development mechanisms. The earliest preserved Paleozoic sediments in all basins (except the Hudson Bay and Moose River basins) are Middle to Late Cambrian. Klein and Hsui (1987) attributed this approximately synchronous basin initiation to the establishment of an extensional regime in the craton, related to the breakup of a supercontinent at the end of the Precambrian. However, Bally (1989) pointed out the lack of significant rifting in the early Paleozoic history of intracratonic basins and questioned the existence of such a supercontinent.

In the case of the Hudson Bay Basin, Quinlan (1987) suggested that initial sedimentation was platformal, and that basin-centred subsidence may not have commenced until the Devonian. Isopach maps (*see* Figures 2 to 11, Map 2582, map case) suggest, although not strongly, that initial subsidence of the Hudson Bay Basin was basin centred. In any case, Paleozoic sedimentation in this basin was initiated at least 70 million years later than in the Michigan Basin.

Many researchers (e.g., Sloss 1988b; Klein 1991) point to the interbasin correlation of Sloss's (1963) sequence boundaries as evidence of cratonic or global tectonic forces controlling synchronous subsidence and uplift. Other workers (e.g., Quinlan 1987; Bally 1989), noting that rates of subsidence do not correlate between basins, suggest that major sequence boundaries may simply reflect eustatic sea level fluctuations superimposed on variably subsiding cratonic elements.

The crust beneath Ontario's intracratonic basins is variable in character and therefore of uncertain significance to Paleozoic basin subsidence. A branch of the 1.1 billion-year-old Midcontinent Rift underlies the Michigan Basin (Fowler and Kuenzi 1978); however, rifting along this structure predates Paleozoic subsidence and sedimentation by 500 to 600 million years and no major rift or graben structures have been identified in the early Paleozoic succession of the Michigan Basin (Bally 1989).

The crust beneath the Hudson Bay Basin is a complex combination of Proterozoic (the Trans–Hudson Orogen) and Archean (the Superior and Hearne provinces) terranes previously assigned to the Churchill Structural Province (Roksandic 1987; Hoffman 1989; *see* Archean Geology of Ontario: Introduction, Thurston, this volume). Roksandic (1987) suggests that reactivation of Precambrian structures beneath this basin during the Silurian and Devonian formed the Central Hudson Bay Arch, a northwest-trending "half-horst" (Sanford 1987).

The Moose River Basin is underlain almost entirely by the Archean Superior Province, which may account for its separation from the Hudson Bay Basin. Other unique features of the Moose River Basin are that its southern boundary is, in part, fault bounded and its southeastern corner is underlain by the Proterozoic Kapuskasing Structural Zone (*see* Sage, this volume).

Gravity anomalies, indicating excess mass in the crust beneath some intracratonic basins, may be targetting igneous bodies related to Precambrian tectonic processes (e.g., Mesoproterozoic rifting). Anomalies have been identified under both the Michigan and Hudson Bay basins (Quinlan 1987; Roksandic 1987).

Seismic evidence also indicates that the crust beneath intracratonic basins is not preferentially thinner, suggesting that lithospheric thinning models such as that proposed by McKenzie (1978) do not apply to these basins, or that subsequent to thinning, the crust was "modified" in some way (Quinlan 1987).

Data from deep exploration wells have been used to create subsidence history curves through the application of backstripping techniques (cf., Miall 1990, p.420–424). Based in part on the general exponential decrease of subsidence history curves, intracratonic basin subsidence has long been attributed to thermal subsidence (Sleep and Snell 1976). Thermal subsidence results from a crustal heating event (e.g., mantle plume or thinned lithosphere), which initially causes uplift and erosion and is followed by subsidence (which generally follows an exponential decay) to thermal equilibrium conditions. However, thermal subsidence is estimated to be too short lived after a heating event (approximately 50 million years) to account for continued subsidence during the entire Paleozoic and there is little evidence for repeated heating events beneath intracratonic basins (Quinlan 1987).

The detail in subsidence history curves depends a great deal on biostratigraphic resolution and on the ability to identify features such as unconformities. The former is typically not good in the predominantly evaporitic successions of intracratonic basins, and the latter depends on the availability of cores from deep wells. Better quality data and assumptions in modelling subsidence have resulted in the recognition of discrete segments in what were previously described as "noisy" curves (Quinlan 1987; Howell and van der Pluijm 1990). Segmented subsidence history curves indicate variable subsidence rates possibly related to different driving processes and/or variable crustal responses to a continuing driving mechanism.

Another type of subsidence proposed to explain intracratonic basin subsidence is extension-related mechanical or faulted subsidence. Faults are known in these basins, but significant graben-type structures are not known, except for the Hudson Bay Basin. In addition, these basins subsided during a dominantly compressive tectonic regime, not during extension.

Models for intracratonic basin subsidence have been proposed which link cratonic tectonism with plate-margin collisional events or orogenic tectonism (e.g., Cloetingh 1988; Sanford et al. 1985; Quinlan and Beaumont 1984; Rabinowicz et al. 1983; DeRito et al. 1983). Quinlan (1987) pointed out that the lithospheric flexure model of Quinlan and Beaumont (1984), in which the development of a foreland basin and associated peripheral bulge is caused by loading of thrust sheets during an orogenic event (Figure

25.23), only modifies the subsidence of an adjacent intracratonic basin, it does not cause it. Quinlan also pointed out that flexural effects would only be realized up to a maximum of 600 km from the basin-forming load, and thus vertical movements of arches and basins farther inboard than the Michigan Basin cannot have been caused by flexural processes.

Rabinowicz et al. (1983) proposed that the plate overriding a subduction zone would be drawn down or tilted, and that this lithospheric tilting could occur up to 1600 km inboard of the plate margin. Quinlan (1987) suggested that the platformal pattern of sedimentation that may have occurred in the Hudson Bay Basin during the Ordovician to Silurian may have been a consequence of lithospheric tilting related to subduction in the East Greenland Orogen. Again, this process does not initiate intracratonic basin subsidence, it may simply modify sedimentation patterns on the craton.

Perhaps the most plausible models for intracratonic basin subsidence involve the horizontal transmission of stresses from plate-margin collisional events into the craton (Quinlan 1987; Cloetingh 1988). Lambeck (1983) and DeRito et al. (1983) presented models which suggest that such stresses are transmitted horizontally until they encounter crustal heterogeneities (e.g., a pre-existing lithospheric curvature) where the stresses are then deflected to induce vertical crustal movements. Such crustal heterogeneities exist beneath the Michigan Basin in the form of the Mesoproterozoic Midcontinent Rift (Fowler and Kuenzi 1978), and beneath the Hudson Bay Basin in the form of structural discontinuities related to the Trans–Hudson Orogen (Roksandic 1987).

Stress propagation models depend on the physical properties of the lithosphere (such as rigidity). Klein (1991) pointed out that the transmission of compressive stresses would be favoured during later Paleozoic orogenic events (i.e., the Alleghanian Orogeny), when the crust was cooler and more rigid. However, Howell and van der Pluijm (1990), recognizing the temporal association of basin-centred subsidence within the Michigan Basin and orogenic events in the Appalachian Orogen, proposed a new mechanism to link these 2 processes. These authors proposed that horizontal stresses, induced in the lithosphere by Appalachian orogenic activity, decreased the viscosity of the lower crust, in effect decoupling the upper crust and mantle. The body of excess mass occurring in the upper crust, as indicated by a gravity anomaly, would then sink into the less viscous lower crust, causing subsidence of the surface (Figure 25.24). Howell and van der Pluijm (1990) further suggested that this process is reactivated during subsequent orogenic events.

Arch uplift, even less well understood than intracratonic basin subsidence, is undoubtedly related to mechanisms driving basin subsidence, either through differential application of these mechanisms, relaxation of driving forces or inverse processes (uplift) acting in interbasinal regions. As mentioned above, the flexure model of Quinlan and Beaumont (1984) does not appear to extend far enough into the craton to explain the activation of any arches except the Algonquin and Findlay arches.

In summary, extension, lithospheric thinning and thermal subsidence may have been significant processes in setting the stage for intracratonic basin development, but these processes probably ceased well before the Paleozoic (Quinlan 1987; Bally 1989). However, these processes may have resulted in the emplacement of an isostatically uncom-

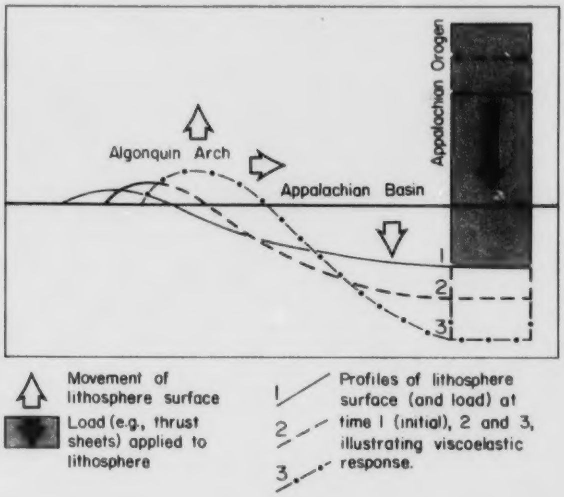

Appalachian Orogen
Algonquin Arch
Appalachian Basin
1
2
3
Movement of lithosphere surface
Load (e.g., thrust sheets) applied to lithosphere
Profiles of lithosphere surface (and load) at time 1 (initial), 2 and 3, illustrating viscoelastic response.
1
2
3

**Figure 25.23.** Schematic cross section illustrating lithospheric flexure model as applied to the Appalachian Basin and Algonquin Arch (*modified from* Quinlan and Beaumont 1984).

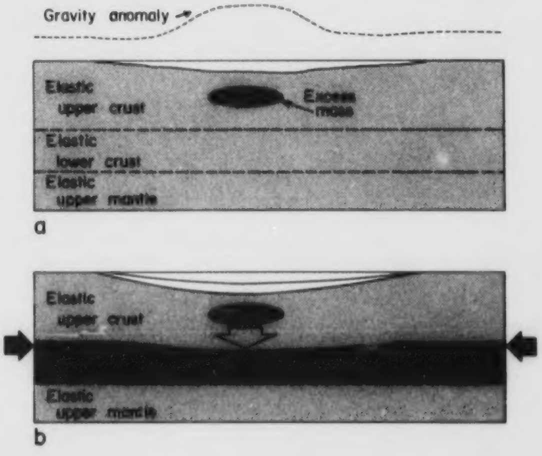


**Figure 25.24.** Schematic lithospheric section, illustrating intracratonic basin subsidence due to crustal weakening induced by compressional tectonism. **a)** Entire crust and upper mantle behave elastically, providing regional flexural support for upper-crustal excess mass (indicated by gravity anomaly). **b)** During peak orogenic activity at plate margin (i.e., compressional regime), elevated intraplate stresses (solid arrows) reduce the effective viscosity of the lower crust. The excess mass can no longer be supported and the upper crust subsides (blank arrow) into the lower crust (*modified from* Howell and van der Pluijm 1990).

pensated mass (igneous body related to deep Proterozoic rifting?), which periodically drove basin subsidence when stresses caused by orogenic activity weakened the lower crust (Howell and van der Pluijm 1990). Flexural effects probably influenced basin and arch movements in near-marginal positions (Quinlan and Beaumont 1984).

## STRUCTURAL EVIDENCE OF TECTONIC ACTIVITY

Intracratonic tectonic activity may have also had a more localized effect on sedimentation. Sanford et al. (1985) postulated a network of fractures forming the boundaries of approximately rectilinear blocks in the Paleozoic strata of southwestern Ontario (Figure 25.25) to explain the occurrences of linear trends in sediment and facies distribution, and linear collapse features related to dissolution of evaporites. The fractures are thought to be the result of brittle deformation of Precambrian basement and overlying sediments in response to cratonic tectonism (i.e., relative vertical movements of basins and arches). The fractures may have developed, in part, along reactivated Precambrian structures (e.g., shear zones). Similar fracture-framework models have been proposed for the Paleozoic of western Pennsylvania (Harper 1989) and Ohio (Coogan 1991).

Faults cutting or underlying Paleozoic and Mesozoic strata in Ontario are shown on the accompanying Bedrock Geology of Ontario and Tectonic Assemblages of Ontario maps (*see* Maps 2541, 2543, 2544, 2575, 2577, 2578, map case). Two classes of faults are shown on these maps: 1) those exposed in surface outcrops, exhibiting a demonstrable vertical offset; and 2) those with no surface expression. The second class includes 2 types: a) those which are interpreted (based on geophysical data) to cut

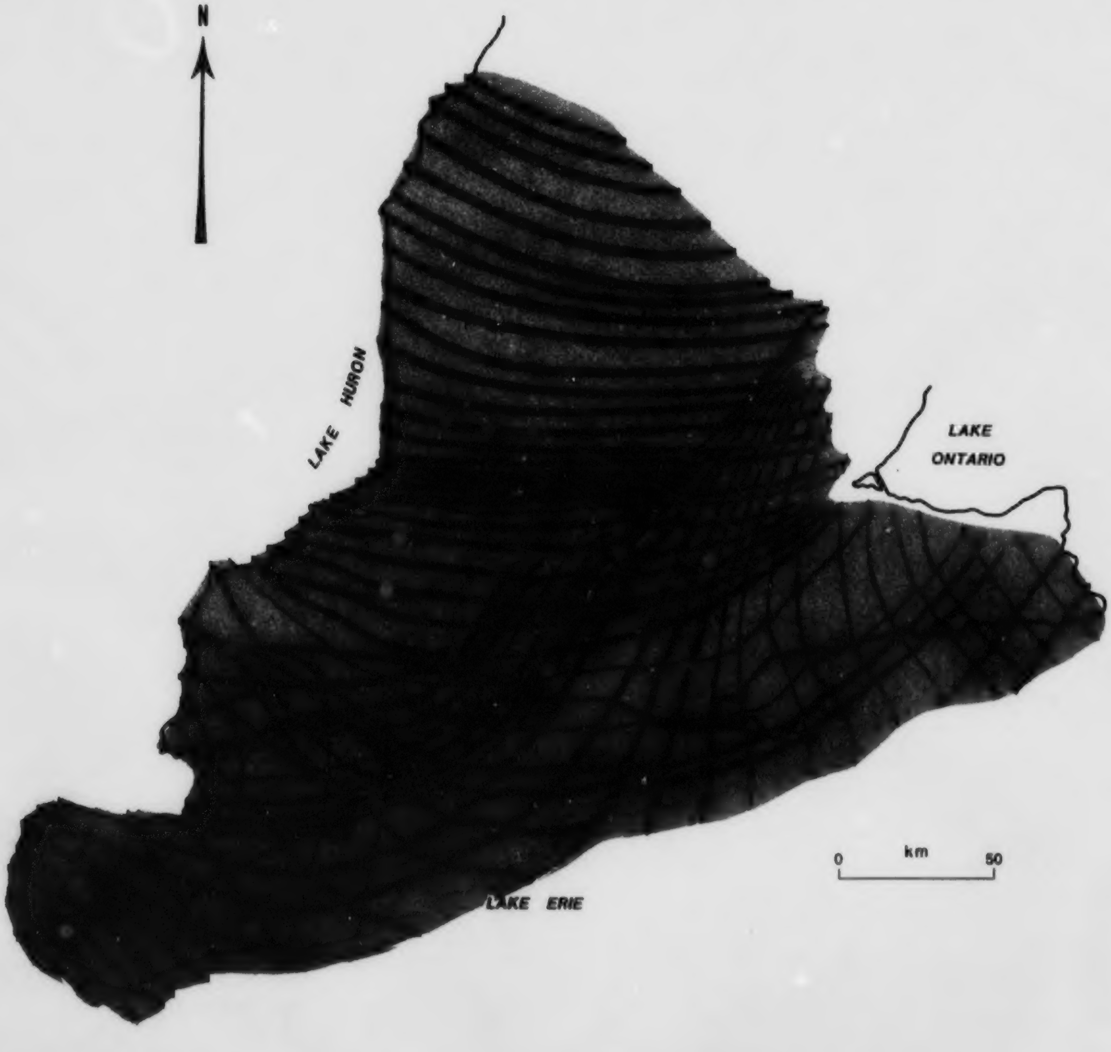


**Figure 25.25.** Conceptual fracture-framework model for southwestern Ontario (*modified from* Sanford et al. 1985).

Precambrian basement, but have no known Paleozoic or Mesozoic displacement; and b) those demonstrably offsetting Paleozoic strata in the subsurface, but with no surface expression. The latter type (b) are only identified in southwestern Ontario where the relative abundance of subsurface information is afforded by hydrocarbon exploration activity. Fault types (a) and (b) are only distinguished on Map 2578 (map case). To date, an extensive fracture framework such as conceptualized by Sanford et al. (1985) has not been identified. This may be attributed, in part, to the extensive cover of Quaternary sediments over most of the area underlain by Paleozoic and Mesozoic strata in the province.

The timing of faulting is not well constrained in Ontario. Faults are documented cutting Ordovician strata in southeastern and south-central Ontario (Williams 1991; *see* Map 2544, map case), and may be responsible, through the dissolution of Silurian salt beds, for controlling deposition of some Devonian facies in southwestern Ontario (Brigham 1971; Sanford et al. 1985; *see* Sequence 7, Johnson et al., this volume). Jurassic kimberlitic dikes cutting Ordovician strata in the Belleville area (Barnett et al. 1984) are slickensided (McFall 1990), suggesting Jurassic or younger movement. Some recent (i.e., historic) seismic activity is reported to be approximately coincident with faults cutting Paleozoic rocks in Ontario or adjacent jurisdictions (Fakundiny et al. 1978; Adams and Basham 1989). Investigations of neotectonic (i.e., post-glacial) features (McFall and Allam 1989; Mohajer and Eyles 1991) also suggest recent reactivation of old structures.

## TECTONIC CONTROLS ON MINERAL DEPOSITS

The Paleozoic and Mesozoic strata of Ontario are hosts for a variety of economic deposits: oil and gas, salt, gypsum, sandstone, shale and carbonates, with diverse uses including brick and tile manufacturing, building and dimension stone, construction aggregate, cement and the chemical industry. Production from the Paleozoic and Mesozoic of Ontario amounted to a value of over $550 million in 1989, or approximately 7.6% of the province's mineral production (B. Greenwell, Ministry of Northern Development and Mines, personal communication, 1991). As discussed in Johnson et al. (this volume), large-scale tectonic forces affected the deposition of sediment types and their related mineral resources. Of these resources, the localization of oil and gas resources appears to be most directly affected by tectonically induced structural controls.

Oil and gas production in Ontario is from units of Cambrian to Devonian age. Reservoirs occur in variety of trap types, including stratigraphic (structurally modified in part) pinch-outs against the Algonquin Arch (Cambrian sandstones), lithostratigraphic pinch-outs (Silurian sandstones and reefs, Devonian carbonates) and structurally controlled dolomitization (Ordovician carbonates). Sanford et al. (1985) suggested how tectonically induced structures are significant in the development of all of these reservoir types.

Subeconomic Mississippi Valley-type lead and zinc sulphide mineralization is well known in the Silurian of southern Ontario (Guillet 1967). This mineralization commonly takes the form of open-space (vugs and fractures) fillings and as replacement of the carbonate matrix (Tworo 1985). Faults or fractures may have been important conduits for mineralizing fluids; however, fissure-related mineralization has only been identified at one locality, the Dundas quarry, near Hamilton (Tworo 1985). Mississippi Valley-type deposits are commonly related to the development of unconformity related secondary porosity (Anderson and Macqueen 1988). Thus, the identification of unconformities and an understanding of their genesis is required in order to evaluate the potential occurrence of such deposits.

The identification of tectonic structures in the Paleozoic and Mesozoic succession will be necessary should these strata be identified in future as potential deep waste-disposal sites.

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# Miscellaneous Thematic

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## Chapter 26

# A U-Pb Geochronological Framework for the Western Superior Province, Ontario

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## Abstract

A compilation of precise U-Pb ages for Archean magmatism in the western Superior Province indicates that the upper crust evolved over a period of more than 500 million years—between about 3200 and 2650 Ma. This evolution included the following: 1) an early stage of intermittent, episodic magmatism between about 3200 and 2800 Ma that appears to be restricted to specific domains; and 2) a main, widespread stage of intensive magmatism between 2770 and 2660 Ma that progressed from early episodes of arc-type magmatism, very similar in style to activity that occurred prior to 2800 Ma, to an orogenic stage characterized by voluminous calc-alkalic magmatism, deposition of turbidite sequences, polyphase folding, thrusting and local alluvial-fluvial sedimentation and alkalic magmatism. These orogenic processes are the combined response to a regime of north-south compression that led to the amalgamation of individual arc-sequences and earlier greenstone segments into a consolidated craton. The timing of the orogenic stage varies diachronously across the Superior Province: it occurred prior to 2710 Ma in the north, between 2710 and 2690 Ma in the centre, and between 2700 and 2680 Ma in the south. The orogenic stage was followed by a protracted maturation of the lower crust that lasted for 50 to 100 million years after formation of the upper crust. This involved complex metamorphism and local plutonism and was apparently responsible for the hydrothermal regimes that formed some gold deposits. The U-Pb ages, together with the available Sr, Nd, Pb and Hf isotopic evidence, indicate that the bulk of the Superior Province was built mainly from juvenile, mantle-derived material with only local involvement of older crustal components. There is little evidence supporting the existence of an extensive, province-wide, early sialic crust.

## INTRODUCTION

The Superior Province has been studied using geochronological methods for the past 40 years. Most of the early work was done using K-Ar and Rb-Sr techniques, and generally yielded ages ranging from about 2700 to 2200 Ma and averaging about 2500 Ma (Stockwell 1982). The K-Ar and Rb-Sr ages often appear to reflect low-grade hydrothermal events and postcrystallization mobility of alkali elements (Beakhouse et al. 1988). In contrast, the earlier U-Pb analyses of zircon and titanite defined ages of about 2600 to 2700 Ma (Tilton and Grünenfelder 1968; Hart and Davis 1969; Steiger and Wasserburg 1969; Krogh and Davis 1971; Catanzaro and Hanson 1971; Peterman et al. 1972). Uranium-lead zircon studies and some Rb-Sr whole rock studies also established that there were gneisses dating as far back as 3.0 Ga (Krogh and Davis 1971; Krogh et al. 1976; Wooden and Goodwin 1980; Clark et al. 1981), and there was some suggestion that older rocks were more abundant in the northern half of the province. Rocks which are over 3.5 billion years in age have been found south of Ontario in the Minnesota River Valley and Watersmeet area of Michigan (Goldich and Wooden 1980; Goldich and Fischer 1986; Peterman et al. 1980). These rocks, however, appear to belong to a separate and older craton that was accreted to the Superior Province *circa* 2.7 Ga or perhaps even later at 1.8 Ga (Sims et al. 1980; Card 1990).

Although the earlier geochronological work was of insufficient precision to resolve individual igneous events within greenstone belts, it was important in establishing a general absolute age for formation of rocks and metamorphism in the Superior Province.

The past decade has seen the widespread application of high precision U-Pb isotope dilution techniques that use low-contamination analytical procedures (Krogh 1973) and the abrasion method (Krogh 1982). This permits the determination of accurate ages with $2\sigma$ errors of about 1 to 5 million years. The U-Pb technique has been applied mainly to zircon, but useful results have also been obtained from accessory titanite, monazite, baddeleyite, rutile, apatite and garnet.

Correct interpretation of ages requires a good understanding of the conditions under which the sampled minerals crystallize. In many igneous rocks, zircons constitute a morphologically uniform population having only one reproducible age. In these cases interpretation is straightforward: the zircon age dates the igneous crystallization of the rock. Complex populations comprising more than one generation of zircons, either as separate grains or as cores and overgrowths, occur in polymetamorphic gneisses or in igneous rocks that contain older zircons inherited from the source of the melt or from incompletely digested wall rock. The ability to precisely analyze small fractions, which may consist of single grains or parts of grains, is a significant advantage in dealing with such rocks. Because of the necessity of obtaining concordant data, only grains free of cracks and alteration can be analyzed, creating the potential for a biased selection that may exclude genetically distinct types of grains if they are less suitable for analysis. In order to avoid similar discriminations, every zircon population is carefully screened and characterized prior to analysis. This allows one to focus the analyses, where possible, on zircons that are representative for all the dominant subgroups. The ages of morphologically different grains may be indistinguishable, implying crystallization under varying physicochemical conditions, but approximately at the same time. Alternatively, they may yield distinct and resolvable ages. In this case, several replicate analyses of one subgroup must show agreement before the age can be trusted.

The interpretation of the significance of an age must take into account the relative abundance of zircons in each subgroup, their morphological characteristics, as well as information concerning the composition and history of a rock and its environment. In rare cases, multiple analyses of zircon from a single population can show differences despite the absence of morphological diversity. This can occur, for example, if zircons contain "hidden" inherited cores. In other cases (e.g., mafic dikes), a rock may only contain xenocrystic or metamorphic zircons and no primary magmatic zircons (Corfu and Andrews 1987; Corfu et al., 1991). Besides the ability to analyze microgram-size samples, the routine careful characterization and selection procedure is one of the key elements necessary for the determination of reliable ages.

The ages obtained from monazite, titanite and garnet can be more difficult to interpret since these minerals are more easily affected by postmagmatic, metamorphic processes that reset the U-Pb systems. The resetting can be due to Pb diffusion with, or without, concomitant recrystallization of the mineral. Often, the younger ages reflect formation of the particular mineral during discrete thermal or hydrothermal events (Corfu and Muir 1989b; Corfu 1988; Wong et al., 1991; Jemielita et al. 1990; Schandl et al. 1990).

This paper presents a broad summary of the ages for magmatism obtained in the Ontario portion of the Superior Province (Figure 26.1) and discusses the main constraints which they impose on Archean development of the Province. We have compiled about 300 ages for magmatic events and about 80 ages for detrital zircons. Most of these ages were obtained at the Jack Satterly Geochronology Laboratory of the Royal Ontario Museum, with additional ages determined at the University of Windsor and the University of Kansas (Turek et al. 1982, 1984, 1988, 1990), and at the Geological Survey of Canada (Sullivan et al. 1985; Mortensen 1987; Percival and Sullivan 1988; Anglin et al. 1989).

Only reliable ages have been considered. All those cases where excessive discordance, poorly resolved inheritance or complex metamorphic overprints complicate the age relationships and yield either minimum ages or ages whose interpretations are unclear have been excluded. These dates have been obtained mainly from zircon and rarely from baddeleyite. Titanite and monazite ages have been used only where they unequivocally record magmatic events. Metamorphic and hydrothermal events that are recorded chiefly by titanite, monazite and rutile are not directly considered in this compilation. Ages for magmatic and metamorphic processes in the lower crust, such as those determined in the Kapuskasing Structural Zone and the Winnipeg River Subprovince, have also not been included in the diagrams. The major temporal features characterizing the evolution of the lower crust are, however, briefly touched on in the final discussion.

Most U-Pb ages considered in the compilation have analytical uncertainties of 1 to 5 million years. They have been assigned to time intervals of 10 million years and are plotted in histograms in Figures 26.2 to 26.7. The compilations present ages for extrusive and intrusive magmatic events. The subdivision into volcanic and plutonic rocks is not always straightforward, since some of the intrusions, for example, dikes and sills, are actually parts of volcanic complexes and were occasionally collected to date volcanic episodes. These distinctions, though important on a local scale, are less critical in the context of the broad comparison carried out in this paper.

Although it is often convenient to summarize geochronological data as a histogram of ages, this may give a biased picture of time development for several reasons. First, geochronological work has generally taken the form of detailed studies in restricted, well-mapped areas, such as economically important parts of greenstone belts. Nonmineralized areas, such as the dominantly metaplutonic terranes, as well as the more remote greenstone belts in the northern part of the Superior Province, have generally been mapped in less detail and have received little geochronological attention. Second, while zircons tend to be abundant in felsic to intermediate, quartz-bearing volcanic rocks, which are often found at middle and upper stratigraphic levels, they are rare in predominantly mafic assemblages such as tholeiitic basalt sequences of greenstone belts, which are often found occupying a basal position in greenstone belts. Therefore, peaks in the age spectrum may, to some extent, reflect a geologically biased sample distribution. However, some dating has recently been performed on all the major greenstone belts of the western Superior Province, and, within the greenstone belts, an attempt has generally been made to cover the most significant assemblages. Thus, despite the above considerations, the data base is now sufficiently large, and regional age patterns sufficiently consistent, that, while future dating and a more extensive coverage may bring some modifications to the present picture, the overall pattern is not likely to change dramatically and can probably be considered as a good approximation of the geochronological relationships in the Superior Province.

## PROVINCE-WIDE AGE DISTRIBUTION

Figure 26.2 shows a compilation of all published ages obtained for magmatic events in the Superior Province of Ontario (*see* Figure 26.1). The main features of the age distribution are: 1) a protracted period between about 3200 and 2800 Ma characterized by sporadic and intermittent magmatism, and 2) a dominant, relatively short-lived period of very intense magmatism between about 2770 and 2660 Ma. The 2 patterns for the ages of volcanism and plutonism are fairly similar. The main difference is a slight, but significant, shift in the timing of the younger activity, with volcanism peaking between 2750 and 2700 Ma, while plutonism was most intensive between 2740 and 2670 Ma with a sharp peak at around 2700 Ma.

## REGIONAL AGE DISTRIBUTIONS

The overall compilation gives a good indication of the main period of crustal development in the Superior Province, but does not reveal regional variations. These are important in evaluating crustal development models, such as the Langford and Morin (1976) model of southward accretion of arcs, which predicts a southward younging of magmatic activity. To examine this possibility, we have subdivided the western Superior Province into 3 separate segments (*see* Figure 26.1): 1) a northern segment which comprises the granite-greenstone Sachigo and Uchi (including the Berens River) subprovinces and the metasedimentary English River Subprovince to the south; 2) a middle segment comprising the gneissic-granitoid Winnipeg River Subprovince, the granite-greenstone Wabigoon Subprovince and the metasedimentary Quetico Subprovince to the south; and 3) a southern segment comprising the granite-greenstone Wawa and Abitibi subprovinces. The Pontiac Subprovince in Quebec forms the southern metasedimentary member of this segment.

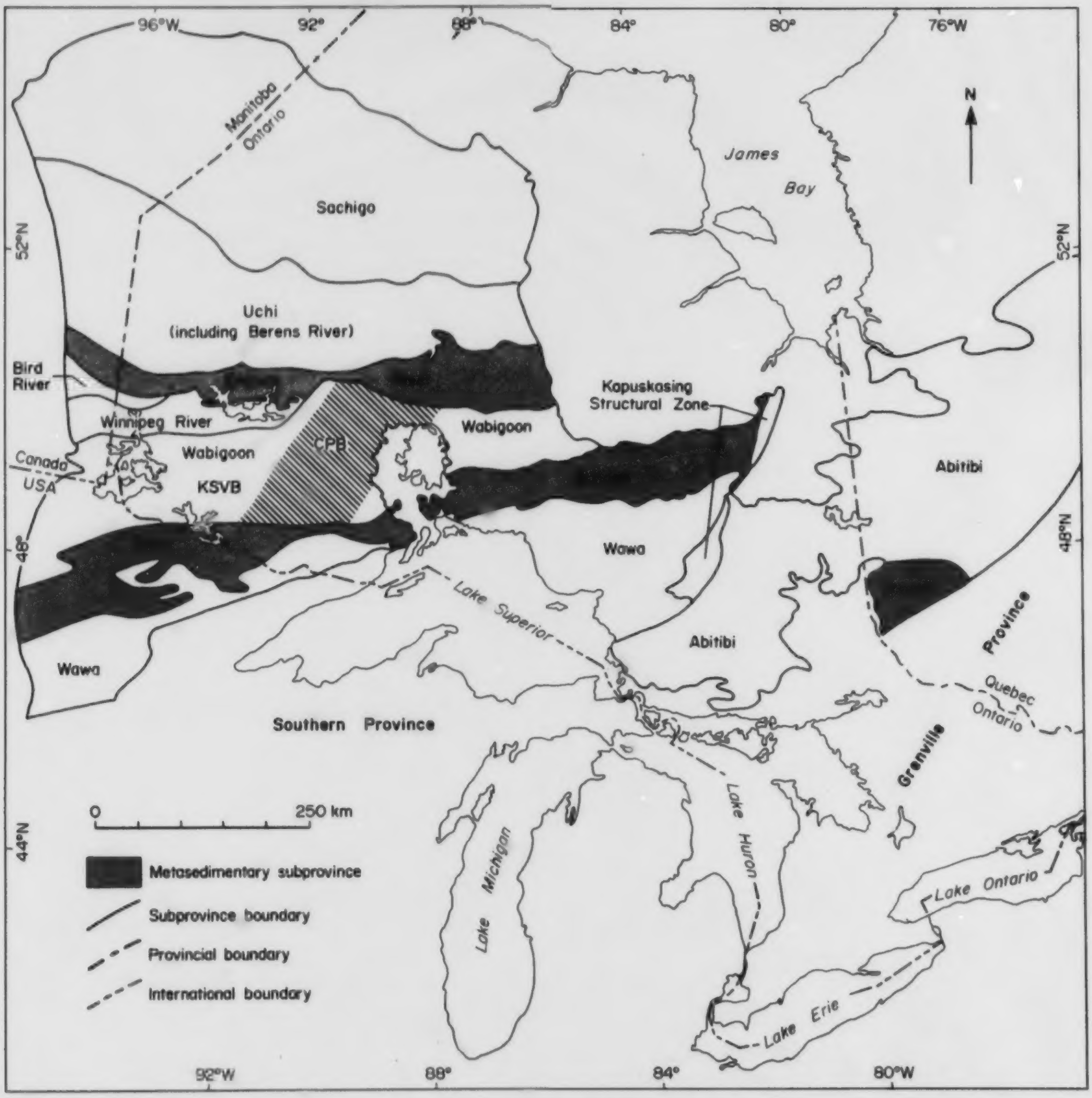
96°W
92°
88°
84°
80°
76°W
Manitoba
Ontario
N
James Bay
Sachigo
52°N
52°N
Uchi (including Berens River)
Bird River
Winnipeg River
Wabigoon
CPB
Wabigoon
Kapuskasing Structural Zone
Abitibi
Canada
USA
KSVB
48°
48°N
Wawa
Lake Superior
Abitibi
Province
Quebec
Ontario
Grenville
Wawa
Southern Province
Lake Huron
0
250 km
44°N
Metasedimentary subprovince
Subprovince boundary
Provincial boundary
International boundary
Lake Michigan
Lake Ontario
Lake Erie
92°W
88°
84°
80°W

**Figure 26.1.** Simplified geological map of the western Superior Province showing the subdivision into subprovinces (*modified from* Card and Ciesielski 1986). The western Wabigoon Subprovince is divided into 2 segments: the Kakagi Lake–Savant Lake volcanic belt (KSVB) and the central plutonic belt (CPB).

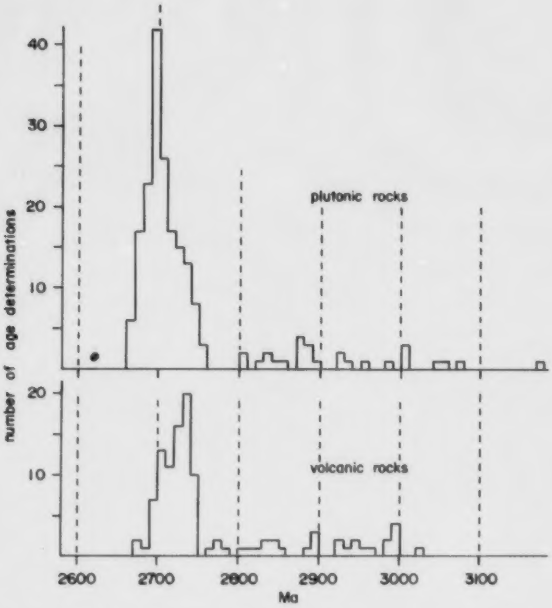

**Figure 26.2.** Histograms of U-Pb ages for plutonism and magmatism in the Ontario portion of the western Superior Province. Most of the compiled ages have uncertainties of 1 to 5 million years and have been assigned to time intervals of 10 million years. Sources of the data are given in the captions to Figures 26.5 to 26.7.

The southern limit of each of these segments consists of a major metasedimentary belt. Williams (1990) and Percival and Williams (1989) have proposed that one of these basins (Quetico Subprovince) may represent an accretionary prism. In this case, the metasedimentary belts would separate individual segments of the Superior Province which may have evolved independently of each other until amalgamation during closure of the intervening oceans.

The comparison for ages of volcanism in Figure 26.3 highlights 2 main features.

1. The earlier period of intermittent volcanism ending at around 2800 Ma was most significant in the northern Superior Province (Nunes and Thurston 1980; Corfu and Wallace 1986; Corfu and Ayres 1991), where it includes nearly half of the age determinations obtained so far. In contrast, volcanism that occurred before 2800 Ma is only recorded in isolated blocks of the middle and southern Superior Province such as in the Wawa Subprovince (Turek et al. 1984) and the central plutonic belt (CPB) in the Wabigoon Subprovince (Davis and Jackson 1988); volcanic components dating from before 2800 Ma are also present in the Winnipeg River Subprovince, but their ages are only constrained by intrusive relationships (Corfu 1988; Beakhouse, this volume).
2. The major period of late volcanism shows a distinctly younger peak in the southern segment relative to the central and northern ones. Although the central and northern segments have similar peaks, the distribution for the central segment is skewed toward younger ages so that it has a younger mean age. Thus, the earliest units from the main period of volcanic activity are approximately coeval in all 3 segments, but the mean age of activity in the regional segments shows a distinct younging from north to south across the Superior Province.

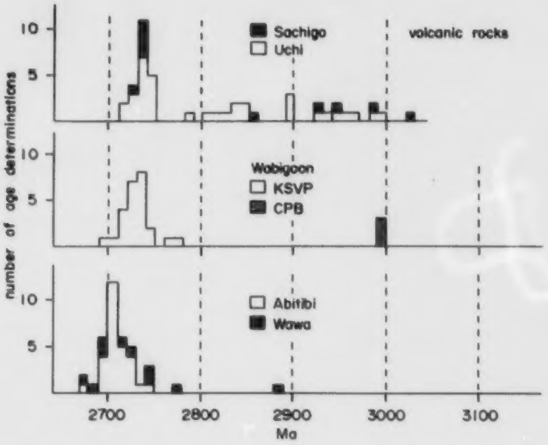

**Figure 26.3.** Comparative histograms with U-Pb ages for volcanic rocks in 3 segments of the Superior Province. **KSVB**—Kakagi Lake–Savant Lake volcanic belt, **CPB**—central plutonic belt.

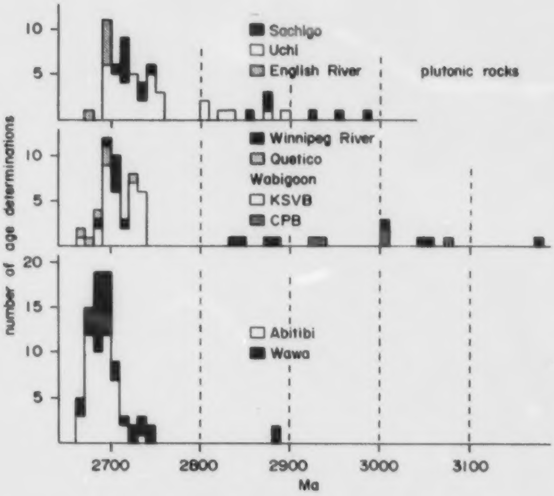

**Figure 26.4.** Comparative histograms with U-Pb ages for plutonic rocks in 3 segments of the Superior Province. **KSVB**—Kakagi Lake–Savant Lake volcanic belt, **CPB**—central plutonic belt.

The ages for plutonism in Figure 26.4 show a picture very similar to that of the ages for volcanism, with both the trend of southward declining of components dating from before 2800 Ma and the gradual southward younging of the late magmatism. The large number of plutonic units emplaced before 2800 Ma in the middle segment is due mainly to the older tonalitic components that occur within the mostly metaplutonic Winnipeg River Subprovince.

The general correspondence between the ages of volcanism and plutonism is evident from the comparisons within individual segments (*see* Figures 26.5 to 26.7). From about 3000 Ma down to, and including, the early part of the 2770 to 2660 Ma peak, the volcanic and plutonic rocks evolved at about the same rate and over comparable time spans. The local occurrence of coeval plutonic and volcanic phases (Davis et al. 1982; Davis and Edwards 1982) indicates that, in many cases, greenstone assemblages evolved as parts of volcano-plutonic complexes. The age distributions which postdate 2770 Ma, however, tend to be broader and have younger average ages for plutonic than for volcanic units. This reflects the fact that, with few exceptions, the youngest generations of plutons are not accompanied by correlative volcanic rocks. These plutons were generally emplaced during a compressional regime that caused folding and thrusting (Chorlton 1990). The calc-alkalic to alkalic volcanic rocks of the Timiskaming group in the southern Abitibi Subprovince (Corfu et al., 1991) and similar units in the Mishibishu (Turek et al. 1990) and the Shebandowan greenstone belts (Corfu and Stott 1986) of the Wawa Subprovince are among the few known, broadly syndeformational or intradeformational volcanic suites.

The lack of significant syntectonic to posttectonic volcanic units could also be due to a biased destruction of the geological record, for example, by selective erosional processes. Evidence for such a bias comes from the study of detrital zircon ages in the Wabigoon and the Quetico subprovinces (Davis et al. 1990). The ages of single detrital zircons from the Quetico sediments (Figure 26.5) show a pronounced peak at around 2710 to 2700 Ma, but there are very few grains whose ages coincide with the peak volcanic activity at 2740 to 2730 Ma in the adjacent source area of the Wabigoon Subprovince. There are, however, petrographic arguments suggesting that these sedimentary basins must have received a significant contribution from volcanic sources (Ojakangas 1985). The mismatch in the dominant age of the detritus and the age of known volcanic sources reinforces independently advanced notions (Ojakangas 1972; Ayres 1983; Ayres and Thurston 1985) that volcanoes which formed coevally with the 2710 to 2700 million-year-old plutons were quickly eroded and redeposited in the adjoining sedimentary basins. This is also supported by the ages of plutons which cut the sediments and which constrain deposition to be within 10 million years of the age of the youngest detrital grain (Davis et al. 1990). The histogram in Figure 26.5 also shows that the older detrital zircon ages in the Quetico and the Coutchiching sediments closely match the distribution of ages for plutonic rocks in the adjacent Wabigoon and Winnipeg River subprovinces. The peak in the young end of the detrital grain age spectrum corresponds to the beginning of regional deformation in these subprovinces (Davis and Smith 1991) and would have been a time of significant uplift and erosion.

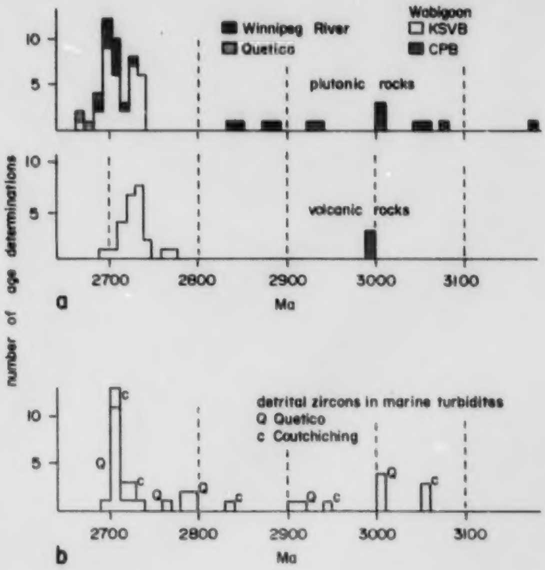

**Figure 26.5 a)** Histograms with U-Pb ages for plutonic and volcanic rocks of the Winnipeg River, Wabigoon and Quetico subprovinces. **KSVB**—Kakagi–Savant Lake volcanic belt, **CPB**—central plutonic belt. **b)** Histogram showing distribution of ages of single detrital zircons for sedimentary rocks of the Quetico Subprovince and of the Coutchiching Group of the Wabigoon Subprovince. Sources for these ages are: Krogh et al. 1976; Davis et al. 1982, 1985, 1988, 1989, 1990; Davis and Trowell 1982; Davis and Edwards 1982, 1986; Davis and Jackson 1988; Davis and Smith 1991; Beakhouse 1983; Morrison et al. 1985; Corfu 1988; Percival and Sullivan 1988; Anglin et al. 1989; Corfu unpublished data; Davis unpublished data.

Ages of detrital zircons from the alluvial-fluvial sedimentary rocks of the Timiskaming group in the southern Abitibi Subprovince (Figure 26.6) show a similar pattern (Corfu et al., 1991). These sequences were deposited on an already consolidated orogen together with emplacement of coeval calc-alkalic to alkalic igneous rocks (Cooke and Moorhouse 1969). The dated detrital zircons yield ages that predominantly match those of the associated igneous rocks with only minor additions from the 2730 to 2690 million-year-old greenstone sequences.

In contrast, zircon detritus in quartz arenites of the Sachigo Subprovince shows only ages dating from before 2900 Ma (Figure 26.7). These mature sedimentary rocks occur locally with stromatolitic marbles, iron formation and komatiites, and have been interpreted as representing platform sequences (Thurston and Chivers 1990). Although their age of deposition is not yet well established, they appear to be associated with the older supracrustal sequences which predate 2800 Ma (Thurston, Osmani and Stone, this volume). Uranium-lead dating of detrital zircons in several of these quartz arenites indicates a provenance of the detritus from sources varying in age between 3080 and 2940 Ma, which corresponds roughly to the age of the oldest volcanic and plutonic (tonalitic to granodioritic) rocks of the greenstone belts in the subprovince. The distribution of detrital zircon ages within individual sandstones, however, tends to be very uniform, with total spreads of less than 20 million years, and, in 2 cases, with only 1 age (±3 million years) population present. Only 1 sample was found to contain an age spread of over 100 million years. These homogeneous populations suggest a very local provenance for the detritus. Therefore, the maturity of the sediments

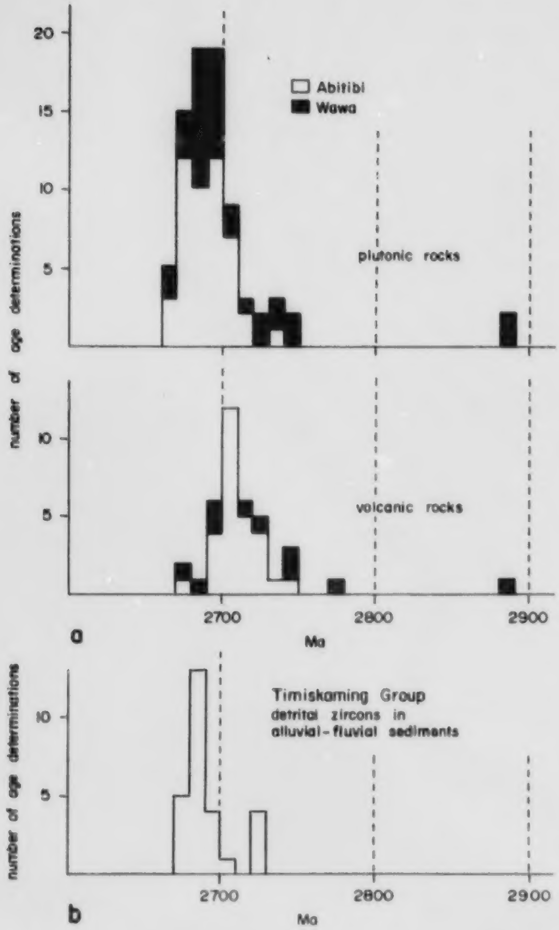

**Figure 26.6 a)** Histogram showing U-Pb ages for plutonic and volcanic rocks of the Abitibi and Wawa subprovinces. **b)** Histogram showing distribution of ages of single detrital zircons for 3 alluvial-fluvial sedimentary rocks of the Timiskaming Group in the southern Abitibi Subprovince. Sources for these ages are: Nunes and Pyke 1981; Krogh and Turek 1982; Krogh et al. 1984; Frarey and Krogh 1986; Turek et al. 1982, 1984, 1988, 1990; Cattell et al. 1984; Sullivan et al. 1985; Mortensen 1987; Corfu and Stott 1986; Corfu and Grunsky 1987; Corfu et al. 1989, 1991; Corfu and Muir 1989a; Wyman and Kerrich 1987; Barrie and Davis 1990; Barrie 1990; Barrie et al. 1990; Wong et al., 1991; Corfu unpublished data.

may be due to local processes such as intense weathering and cannot be used as evidence for the existence of an extensive older cratonic source. The fact that no younger detrital components have so far been found, despite the abundance of volcanic and plutonic units younger than 2900 Ma, supports very early depositional ages for these quartz arenites.

In summary, both the old quartz arenites and the alluvial-fluvial Timiskaming-type sediments appear to have received their detritus from immediately surrounding sources. The turbiditic sediments within the large metasedimentary subprovinces had sources with a much wider range of ages. For the Quetico Subprovince, this is generally reflected by the regional age distribution of surrounding rocks, but there appears to be an underrepresentation of detrital zircons coeval with exposed volcanic rocks. As with the other 2 types of sediment, a significant proportion of the detrital zircons came from igneous rocks which were emplaced shortly before deposition of the sediments.

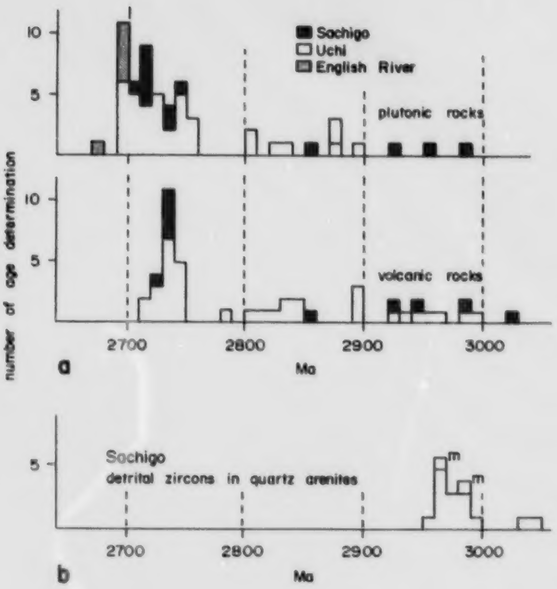

**Figure 26.7 a)** Histograms with U-Pb ages for plutonic and volcanic rocks of the Sachigo, Uchi and English River subprovinces. The few ages available from the Berens River Subprovince have been included in the histogram for the Uchi Subprovince. **KSVB**—Kakagi Lake–Savant Lake volcanic belt, **CPB**—central plutonic belt. **b)** Histogram showing distribution of ages of single detrital zircons in several quartz arenites of the Sachigo Subprovince. **m** denotes the age of uniform detrital zircon populations found in two of the sandstones; each age is based on multiple analyses of single grains and/or multigrain fractions. Sources of these data are: Nunes and Thurston 1980; Corfu and Ayres 1984; Corfu et al. 1985; Corfu and Wood 1986; Corfu and Wallace 1986; Corfu and Andrews 1987; Noble 1989; Schärer 1989; Corfu and Stott 1989; Corfu and Ayres 1991; Corfu unpublished data; Davis unpublished data.

## DISCUSSION

Uranium-lead geochronology indicates that the evolution of the western Superior Province was characterized by 1) an early period of intermittent magmatism lasting from about 3200 to about 2800 Ma, of which the rocks older than 3.0 billion years are mostly plutonic, and 2) major magmatism between 2770 and 2660 Ma.

The magmatism that occurred prior to 2800 Ma was relatively widespread in the northern Superior Province, but it only formed isolated segments of the southern Superior Province. In general, these early volcano-plutonic complexes appear to have remained relatively isolated until they were amalgamated with the younger units during the major deformational events *circa* 2700 Ma. There is some sparse evidence supporting events causing deformation and metamorphism prior to 2800 Ma, for example, in the Eyapamikama Lake (Sachigo Subprovince; Davis, unpublished data) and Lumby Lake (Wabigoon Subprovince; Davis and

Jackson 1988) greenstone belts, and in the Winnipeg River Subprovince (Beakhouse 1983). These early orogenic pulses remain poorly understood. They may represent local collisions between island arcs, or between island arcs and microcontinents. Although evidence on the extent of older crust is somewhat obscured by the intensity of later magmatic and orogenic activity, there is little indication that a large, consolidated, continental crust, on the scale of the province, existed 2900 to 2800 Ma. The mature quartz arenites that are present on top of the earliest greenstone sequences would be compatible with the existence of a stable continental environment (Thurston and Chivers 1990), but the detrital zircons in these rocks point to very local provenances with none of the complex source characteristics shown, for example, by the mature quartzites at Mt. Narryer and Jack Hills in the northwestern Yilgarn craton in Australia (Froude et al. 1983; Compston and Pidgeon 1986). The lack of substantial older crust in the Superior Province is also reflected by the positive $\varepsilon_{Hf}$-values for detrital zircons in quartz arenites and in turbidites of the Superior Province (Stevenson and Patchett 1990), and by the abundance of "young" (ca. 2700 Ma) ages for detrital zircons in the Pontiac and Quetico Subprovince turbidites (Gariépy et al. 1984; Davis et al. 1990; Davis, unpublished data).

The major period of magmatism in the Superior Province occurred between 2770 and 2660 Ma (*see* Figure 26.2). Some areas, such as the southern Abitibi Subprovince, were formed almost entirely during these late events (Corfu et al. 1989, 1991 and references therein). Other areas, such as the Favourable Lake greenstone belt in the Sachigo Subprovince, experienced the formation of new volcanic and plutonic rocks and their coalescence with crustal segments which predate 2800 Ma (Corfu and Ayres 1991). The 2770 to 2660 million-year-old magmatism can be subdivided into a pre-orogenic stage, in the approximate age range 2770 to 2700 Ma, and a subsequent orogenic stage. As explained below, the beginning of the orogenic stage appears to young diachronously from north to south across the Superior Province.

The pre-orogenic stage mainly involved the development of volcano-plutonic complexes in an ocean setting (e.g., southern Abitibi Subprovince, Kakagi Lake–Savant Lake volcanic belt in the western Wabigoon Subprovince), or locally on older arc sequences (e.g., Favourable Lake in the Sachigo Subprovince, Pickle Lake in the Uchi Subprovince). This stage is thus essentially indistinguishable in style from the magmatic phases which predate 2800 Ma. The major difference is one of volume, the 2770 to 2700 million-year-old pulse of pre-orogenic magmatism was far more intense than the sporadic and localized earlier activity.

The orogenic stage was characterized by a number of distinct geologic processes: 1) deposition of turbidite sequences, the apparent equivalent of flysch in modern orogenic belts, which are now represented by the metasedimentary assemblages within, and adjacent to, greenstone belts; 2) emplacement of calc-alkalic plutonic sequences with diorites, tonalites and granodiorites, including a more potassic suite of granitoid rocks which is extremely rare in the older suites; 3) coalescence of the various greenstone segments into a fold-thrust belt involving the juxtaposition of individual sequences by thrusting and open upright folding; and 4) transpressive deformation, localized in the later stages along faults. Synorogenic volcanic rocks are relatively rare, but, as previously discussed, many of these rocks may have been completely eroded and deposited in the adjacent turbidite basins.

Although involvement of the crust that predates 2800 Ma in the generation of the pre-orogenic 2770 to 2700 million-year-old magmatic suites can be locally observed, most of these rocks were produced from relatively juvenile material. This is indicated by U-Pb and other isotope systematics.

The occurrence of inherited zircons can be an important indicator of the sources and the tectonic setting of magmas. Inheritance is, for example, widespread in granitic melts generated in continental settings (Williams et al. 1990; Schärer and Allègre 1983; Dunning et al. 1989; Palmer and Davis 1987), but it is virtually absent from rocks, such as ophiolites, formed in oceanic settings (Dunning and Krogh 1985; Dunning and Pedersen 1988). In the western Superior Province, xenocrystic zircons are a quite common occurrence in late to postorogenic igneous rocks, reflecting their generation from the melting of older, lithologically evolved units and/or local contamination during emplacement (Corfu et al. 1989, 1991; Corfu and Muir 1989a, Davis and Smith 1991). In these cases, the ages of the xenocrystic zircons reflect crudely the spectrum of ages of the rocks in the area. In contrast, inherited zircons are extremely rare in pre-orogenic arc sequences (*see* references in captions for Figures 26.5 to 26.7). Exceptions include late arc sequences emplaced in greenstone belts that already contain abundant older assemblages such as at North Spirit Lake and Favourable Lake in the Sachigo Subprovince (Corfu and Wood 1986; Corfu and Ayres 1991). These contrasting relationships cannot be explained by sample biasing during selection, and suggest that much of the western Superior Province evolved in an oceanic setting prior to the final amalgamation and consolidation at around 2.7 Ga.

The general absence of significantly older evolved crust is shown especially by the isotopic signatures of Sr, Pb, Nd, and Hf that indicate derivation largely from a depleted mantle source or short-lived mafic crust derived from it ($^{87}Sr/^{86}Sr = 0.7005$ to 0.7010, $^{207}Pb/^{204}Pb = 14.3$ to 14.4, $\varepsilon_{Nd} = +3$ to +2, $\varepsilon_{Hf} = +6$ to +4). Rocks emplaced during the orogenic stage, in some cases, show evidence for contributions from relatively short-lived crustal sources ($^{87}Sr/^{86}Sr = <0.703$, $^{207}Pb/^{204}Pb = <14.6$, $\varepsilon_{Nd} = \geq 0$, $\varepsilon_{Hf} = >3$) (Peterman et al. 1972; Hart and Brooks 1977; Patchett et al. 1981; Turek et al. 1982; Zindler 1982; Beakhouse 1983; Franklin et al. 1983; Tilton 1983; Basu et al. 1984; Dupré et al. 1984; Cattell et al. 1984; Gariépy and Allègre 1985; Ashwal et al. 1985; Morrison et al. 1985; Machado et al. 1986; Shirey and Hanson 1986; Maas et al. 1986; Smith et al. 1987; Jacobsen and Pimentel-Klose 1988; Walker et al. 1988; Corfu and Muir 1989b; Noble 1989; Barrie 1990; Corfu and Noble 1990). Elevated $^{87}Sr/^{86}Sr$ (0.703 to 0.705) and $^{207}Pb/^{204}Pb$ (14.4 to 14.9) and lower $\varepsilon_{Nd}$ (<0) and $\varepsilon_{Hf}$ (<3) are only seen in areas such as the Winnipeg River Subprovince, the Michipicoten greenstone belt and portions of the Uchi Subprovince, which are known to contain some of the oldest crust in the Superior Province (Wooden and Goodwin 1980; Beakhouse 1983; Smith et al. 1987; Thorpe et al. 1987; Corfu 1988; Noble 1989; Dickin et al. 1990).

The isotopic signatures tend to correspond to the general pattern defined by the U-Pb ages and support the notion that the age distribution in Figure 26.2 reflects, in a general way, the real growth rates of the Superior Province. Therefore, the dominant peak of magmatism at 2770 to 2660 Ma also corresponds to the major period of crust formation.

The orogenic stage is related to a common north-south directed compressional regime as shown by the sequence of geological events recorded throughout the Superior Province. Evidence for major thrust displacement is found, for example, in the Favourable Lake greenstone belt in the Sachigo Subprovince (*see* Figure 26.1). The belt comprises 5 assemblages that formed between 2925 and 2725 Ma, but which are juxtaposed stratigraphically out of sequence along at least 3 thrust faults (Corfu and Ayres 1991; Ayres and Corfu 1991). The youngest, a 2725 million-year-old group, contains a thick assemblage of alluvial-fluvial and turbidite sedimentary rocks that have been overthrust by older sequences prior to folding and intrusion of late tectonic 2711 million-year-old plutons.

Evidence for late thrusting is also found in the Kakagi Lake–Savant Lake volcanic belt of the Wabigoon Subprovince to the south (*see* Figure 26.1). Here, 2750 to 2720 million-year-old volcanic units structurally overlie *circa* 2700 million-year-old alluvial-fluvial and turbiditic sedimentary rocks that form a widespread unconformity with underlying basaltic rocks which may be of ocean floor affinity (Davis et al. 1988). An early subhorizontal fabric within the sediments (Chorlton 1990) may reflect deformation during tectonic juxtaposition of the older units on top of the younger ones. Thrusting was probably active during the emplacement of the earliest, 2710 million-year-old syntectonic plutons (Davis and Smith 1991); it affected the *circa* 2700 million-year-old sedimentary sequences and was apparently concluded by the time late plutons were emplaced at 2696 Ma. Deposition of the turbidite sequences in the Quetico Subprovince to the south was also related to this orogenic phase which resulted in erosion of significant amounts of the essentially coeval, synorogenic magmatic rocks (Figure 26.5; *see also* Davis et al. 1990).

The southern Abitibi Subprovince shows a similar pattern of events, including the overlapping deposition of abundant turbidite sequences, calc-alkalic magmatism, thrusting and folding (Corfu et al., 1991). In addition, the southern Abitibi Subprovince experienced a very pronounced late orogenic period of alkalic volcanism and plutonism and associated alluvial-fluvial sedimentation (Cooke and Moorhouse 1969; Hyde 1980). These suites, collectively known as the Timiskaming group, were themselves affected by local thrusting and polyphase folding (Hodgson and Hamilton 1989).

Thus, a few dominant processes such as thrusting, folding, deposition of turbidites, calk-alkalic plutonism and local Timiskaming-type sedimentation and alkalic magmatism accompanied orogeny in most greenstone belts. The timing of the orogenic stage shows a systematic younging from north to south across the Superior Province: it occurred prior to 2710 Ma in the north, between 2710 and 2690 Ma in the centre, and between 2700 and 2680 Ma in the south (*see* Figure 26.4). This presumably reflects the progressive docking of individual greenstone and crustal segments.

Most of the geochronological work in the Superior Province has been directed toward understanding the igneous and structural development of upper crustal rocks, particularly in the greenstone belts. Age results from rocks which have undergone medium to high grades of metamorphism in the lower crust are often more difficult to interpret. Because of this uncertainty, and the sparse amount of published data, only a general discussion will be given of metamorphic ages.

Lower crustal evolution can only be studied in relatively restricted areas, such as the Kapuskasing Structural Zone and the Pikwitonei Subprovince (in Manitoba) where post–Archean faulting has exposed deep crustal levels. In both these areas, which are located almost at opposite ends of the Superior Province, metamorphic zircon crystallization ages are recorded in the lower crust for about 50 to 100 million years following the apparent completion of magmatism and tectonism at high crustal levels (Percival and Krogh 1983; Heaman et al. 1986; Corfu 1987; Krogh et al. 1988; Mezger et al. 1989, 1990). There is also an early phase of metamorphism at around 2700 to 2690 Ma that is present only in the north (Heaman et al. 1986; Mezger et al. 1989). The lower crustal processes include polyphase metamorphism involving both dehydration and rehydration reactions, some plutonism and the development of subhorizontal structures (Percival and Card 1985). The pattern of metamorphic mineral crystallization ages seems to require progressively later additions of heat and fluids to cause metamorphism and local melting at deeper crustal levels (Krogh et al. 1988; Krogh 1990). The cause of this activity may involve magmatic or tectonic underplating or heating associated with crustal extension and/or lithospheric delamination accompanied by rising mantle isotherms.

The high-level response to the lower crustal events appears to be recorded in crustal-scale, steeply dipping shear zones, where hydrothermal alteration led to the development of secondary mineral assemblages including titanite, rutile and monazite that yield U-Pb ages varying from 2650 to 2600 Ma (Corfu and Muir 1989b; Jemielita et al. 1990; Schandl et al. 1990; Wong et al., 1991). Other cogenetic vein minerals such as micas and scheelite provide comparable Ar-Ar and Sm-Nd ages (Hanes et al. 1989; Masliwec et al. 1985, 1986; Anglin 1990). This late activity in the high-level shear zones is of great economic significance because of its association with gold mineralization (Jemielita et al. 1990; Wong et al., 1991).

## CONCLUSIONS

A compilation of precise U-Pb ages for magmatism, obtained mainly from zircons during the past decade, highlights some of the fundamental features that characterize the tectonic evolution of the upper crust of the Superior Province in Ontario.

The ages span over 500 million years and delineate an early period of about 3200 to 2800 Ma, characterized by intermittent episodes of magmatism, and a late period of about 2770 to 2660 Ma, characterized by very intense magmatism and tectonism. The late period can be subdivided into 2 phases: 1) an earlier phase of arc-type volcanism and sodic plutonism that is qualitatively comparable to the evolution of the older greenstone components prior to 2800 Ma; and 2) an orogenic phase that involved thrusting, folding, voluminous sedimentation and emplacement of calc- alkalic to alkalic plutons and batholiths. The deformational event was the result of north-south compression and probably lasted at least 30 million years. The initiation of this compressive event shows a distinct southward younging across the Superior Province and a parallel decrease in the age of the youngest pre-orogenic arc complexes. This may be attributable to progressive docking of arc and microcontinental fragments. A north-south decrease in components which date from before 2800 Ma is also evident in the western Superior Province.

The age relationships among, and within, distinct groups of greenstone belts across the Superior Province, and the distribution of the earliest crustal segments, are consistent with growth as a result of convergence and subduction processes and support the notion that the Archean crust formed by mechanisms comparable to those of modern plate-tectonic settings.

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## Chapter 27

# Regional Geochemical Mapping

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## Abstract

**Geochemical maps are now an essential component of the mineral resource appraisal process as it is applied by government agencies worldwide. For this reason, regional geochemical maps are prepared by the Ontario Geological Survey (OGS) for areas of the province considered to have particularly high mineral potential. Other components of OGS mineral resource appraisal activity include: 1) geological; 2) geophysical; 3) Quaternary geology; and 4) remote sensing maps prepared at a regional scale. A digital data base designed for use with a Geographical Information System (GIS) may employ all of these maps as components.**

**Part I of this chapter shows how regional geochemical maps are designed in relation to the other GIS components. Part II describes the theory behind the methodology used for the preparation of OGS regional geochemical maps. Examples of 3 OGS approaches to regional geochemical mapping are described in Part III of the chapter. Part IV outlines some of the research and development on which the different approaches were based. Part V includes: 1) a general discussion; 2) reference to some recent developments in geochemical mapping; and 3) some conclusions. The conclusions concern the future of OGS geochemical mapping, the need for further regional geochemical mapping research, and the possibility of the preparation of a geochemical map of Ontario.**

## PART I: OVERVIEW

### An Introduction to Regional Geochemical Mapping in Ontario

In 1893, Willet Miller, a geology professor from Queen's University, completed what is usually considered to be the first successful geochemical prospecting in Ontario. Miller did so by locating economic corundum deposits in drift-covered areas of southeastern Ontario. According to Tyrrell (1926, p.104):

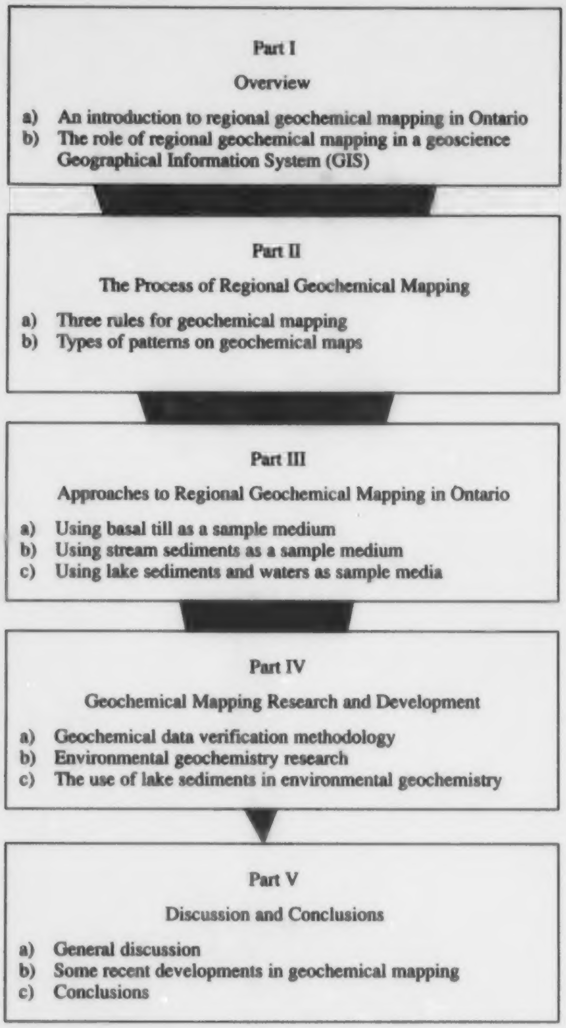
> He [Miller] worked out a new method in the field of discovery. The rock exposures of corundum are not numerous in eastern Ontario. When he had discovered the direction in which the corundum-bearing rock seemed to be trending, he would travel southward, and then turn back north, watching for boulders of corundum bearing rock in the glacial drift. In other words, he came upstream against the glacial flow, and where he found drift boulders he was usually close to the source of supply.

Miller's corundum prospecting sounded a keynote for successful predictive mineral resource appraisal activity in Ontario. Prior to the 1960s, in the Ontario Geological Survey (OGS), this keynote was sounded almost exclusively by geological mapping (Groen 1978). Since 1960, geological mapping for mineral resource appraisal by the OGS has been augmented by: 1) geophysical mapping; 2) Quaternary geological mapping; 3) geochemical mapping; and most recently by 4) remote sensing (Groen 1978, 1987).

It was estimated recently (M. Cherry, Ontario Geological Survey, personal communication, 1990) that an area of 147 000 km$^2$ of Ontario has a high potential for mineral deposits. By 1990, this entire area had been mapped geologically by the OGS at least once and 59% of the area had been mapped geophysically by the OGS. In 1990, regional geochemical mapping of the province was still in its infancy, and only 5.9% (or 5 000 km$^2$) of the high mineral potential area had been mapped by this method. This relatively small area explains: 1) why this type of mapping is seldom mentioned elsewhere in this volume; and 2) the need to stress methodology and mapping results equally, in this introduction to the subject.

During the 1980s, mineral resource appraisal methodologies applied by government agencies passed through a revolution in technology worldwide. The revolution came about with the general introduction of the Geographic Information Systems (GIS) approach. This involves image processing of geographical and geoscience data simultaneously by computer (Fortescue 1990b; Plant et al. 1984; Ryghaug and Green 1988; Wright et al. 1988 and many others).

At the OGS, the gradual introduction of GIS methodologies (combining data sets from geological, geophysical and geochemical mapping) has struck a fresh keynote in the mineral resource appraisal process. In particular, this development has been accompanied by a rethinking of the philosophy for geochemical mapping in the province. Until the 1980s, the general aim of OGS geochemical surveys was to locate geochemical anomalies which could then be explained in terms of the presence of economic mineral deposits. Since 1980, regional geochemical mapping methodologies based on variations of standard methods on 1) basal till, 2) stream sediments and 3) lake sediments, have been developed by the OGS. These methodologies are designed to map geochemical province patterns in addition to geochemical anomalies.



Part I

Overview

a) An introduction to regional geochemical mapping in Ontario
b) The role of regional geochemical mapping in a geoscience Geographical Information System (GIS)

Part II

The Process of Regional Geochemical Mapping

a) Three rules for geochemical mapping
b) Types of patterns on geochemical maps

Part III

Approaches to Regional Geochemical Mapping in Ontario

a) Using basal till as a sample medium
b) Using stream sediments as a sample medium
c) Using lake sediments and waters as sample media

Part IV

Geochemical Mapping Research and Development

a) Geochemical data verification methodology
b) Environmental geochemistry research
c) The use of lake sediments in environmental geochemistry

Part V

Discussion and Conclusions

a) General discussion
b) Some recent developments in geochemical mapping
c) Conclusions

**Figure 27.1.** Flow diagram of the organization of the chapter.

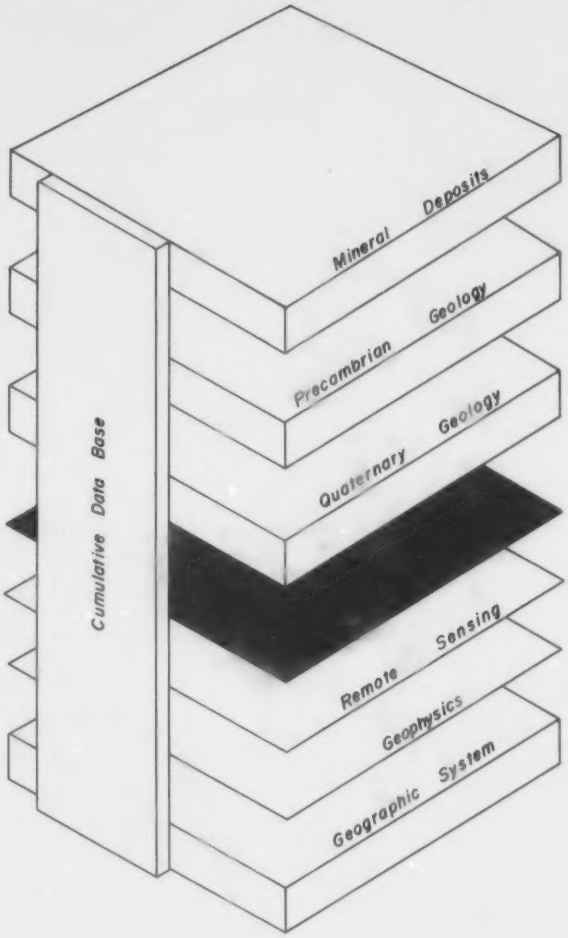


**Figure 27.2.** Block diagram showing relationships among map components of a hypothetical, Ontario Geological Survey regional level, mineral resource appraisal program (*modified from* Fortescue and Vida 1990a).

A flow diagram (Figure 27.1) describes the organization of this chapter in 5 parts.

## The Role of Regional Geochemical Mapping in a Geoscience Geographical Information System

The current series of OGS regional geochemical maps are designed for the use of prospectors, explorationists and users of a Geographic Information System (GIS). The role of these maps in a GIS environment is clarified by a general conceptual model (Figure 27.2). In this proposed model, the ideal sequencing of geoscience data collection for future projects would commence at the bottom of the diagram. Data collection would be accomplished in 3 steps: 1) the digital compilation of all geoscience map data for the area of interest at the start of the project; 2) the completion of new geophysical, geochemical and remote sensing maps of the area based on the digital data base established in Step 1; and 3) the preparation of new surficial, geological and metallogenic maps. The new maps prepared during step 2 would be released to the public soon after their preparation and be made available to OGS field geologists involved in step 3. Incidentally, the cost per km² of map coverage in step 3 is usually ten times greater than the map coverage in step 2.

Consideration of this model has affected the current OGS regional geochemical mapping program in several ways. For example: 1) the scientific tenor of the OGS regional geochemical maps should be comparable to other GIS-oriented regional geoscience data; 2) the time required to produce OGS geochemical maps should be less than, or equal to, that required for OGS geophysical maps; and 3) in order to ensure a uniform standard province-wide, the OGS regional geochemical mapping methodology must be fully standardized.

# PART II: THE PROCESS OF REGIONAL GEOCHEMICAL MAPPING

## Three Rules for Regional Geochemical Mapping

Modern geochemical mapping must follow 3 rules in order to be effective for mineral resource appraisal purposes.

1. The number and distribution of sample collection points must be uniform within the map area.
2. The sampling medium must be uniform throughout the map area.
3. The mapping methodology must be standardized.

Each of these rules will now be discussed in relation to the OGS regional geochemical mapping experience.

### RULE 1: UNIFORM SAMPLE DENSITY

A modern geochemical map is based on sample points distributed uniformly across the mapped area; the number of sample points must be sufficient to describe the principal geochemical patterns in the area. Lake sediment regional geochemical mapping by the OGS relies on a uniform average sample density of 1 sample per km$^2$. The British Geological Survey (BGS) utilizes the same sample density (using stream sediments) for GIS-oriented regional geochemical mapping (Plant et al. 1990).

In Ontario, geochemical maps are currently produced at 2 sample densities with a third density under consideration (Table 27.1).

#### Reconnaissance Level Mapping

Since the mid–1970s, reconnaissance geochemical maps (prepared by the Geological Survey of Canada (GSC) under joint agreements with the OGS) have covered large areas of Ontario (Friske 1985). These maps are of general interest to the mineral exploration industry because low sample density, reconnaissance geochemical maps are now considered to be not detailed enough for use in a GIS, in areas of high mineral potential.

#### Regional Level Mapping

Regional level geochemical mapping was developed in Ontario largely as a result of the findings of a report on this subject (Fortescue et al. 1980). This report was critical of the use of traditional reconnaissance geochemical mapping for: 1) mapping areas of Ontario with high mineral potential; and 2) collecting base line data for environmental geochemistry.

Fortescue and Stahl (1987b) provided a comparison of the coverage of regional and reconnaissance geochemical mapping. The Wart Lake area is located in the Batchawana greenstone belt, some 50 km northeast of Sault Ste. Marie. The southern part of the Wart Lake area is relatively simple granitic terrane. Farther to the north, in the Batchawana greenstone belt, the geology is more complex (Figure 27.3). Fortescue and Stahl (1987b) provided 2 maps of the geology of the Wart Lake area. The dots in Figure 27.3b indicate the reconnaissance level geochemical sample density, while the dots in Figure 27.3c indicate those for the regional level. The distribution of sample points is uniform in Figures 27.3b and 27.3c. On the regional map, the sample intensity is considerably greater. The increased sample intensity facilitates the recognition of geochemical patterns in areas of complex geology (*see* Figure 27.3c).

#### Low Density Sampling Mapping

In 1983, the preparation of a geochemical map of Ontario, based on reconnaissance level sampling, was considered and rejected as being impractical (Fortescue 1983b). The preparation of such a geochemical map is now considered feasible based on the recent Scandinavian experience

**Table 27.1.** Three types of geochemical mapping programs suitable for application in Ontario.

| | | **Area (km$^2$) per sample** |
|---|---|---|
| 1. | OGS Regional Geochemical Maps (lake sediments and lakewaters) | 1 |
| 2. | GSC–OGS Reconnaissance Geochemical Surveys (lake sediments and lakewaters) | 14 |
| 3. | OGS Low Density Geochemical Surveys (overbank samples and lakewaters) | 500 |

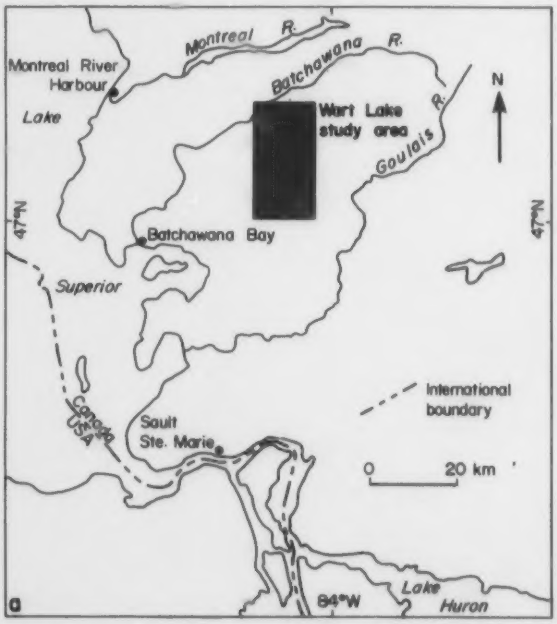


**Figure 27.3a.** Geological maps of the Wart Lake area showing lake sediment sample point locations: generalized map showing the Wart Lake study area (*modified from* Fortescue and Stahl 1987c).

described below. The map would be based on low density sampling which was recently developed in Norway (Ottesen et al. 1989) and Finland (Koljonen et al. 1989). Geochemical surveys of this type are based on a sample density of 1 sample per 500 km². Geochemical surveys of this type are based on: 1) overbank samples (Ottesen et al. 1989); 2) till samples (Koljonen et al. 1989); or 3) water samples (Eden 1990). A small-scale experiment involving overbank sampling in Ontario is described in the section entitled Geochemical Map of Ontario (Fortescue and Bolviken 1990).

## RULE 2: UNIFORM SAMPLE MATERIAL

Effective geochemical surveys are always based on a sample medium which is common over the area selected for mapping. Unfortunately, there is no single sample medium suitable for geochemical mapping of the entire province of Ontario at the regional level. Therefore, the OGS has developed a series of regional geochemical mapping methodologies (based on different sample media) for use in various regions of Ontario. In the 1980s, 3 approaches to regional geochemical mapping were developed. These approaches were based on: 1) basal till (for use in areas of deep overburden, e.g., in the Kirkland Lake area); 2) stream sediments (for use in the limestone terrains of southwestern Ontario); and 3) lake sediments (for use in the southern Canadian Shield where marl lakes are absent).

Within a given region, the general geology of the bedrock and/or the nature of the Quaternary cover govern the choice of sample media for regional geochemical mapping. Figure 27.4 is a series of general conceptual landscape models drawn to illustrate this phenomenon. For example,

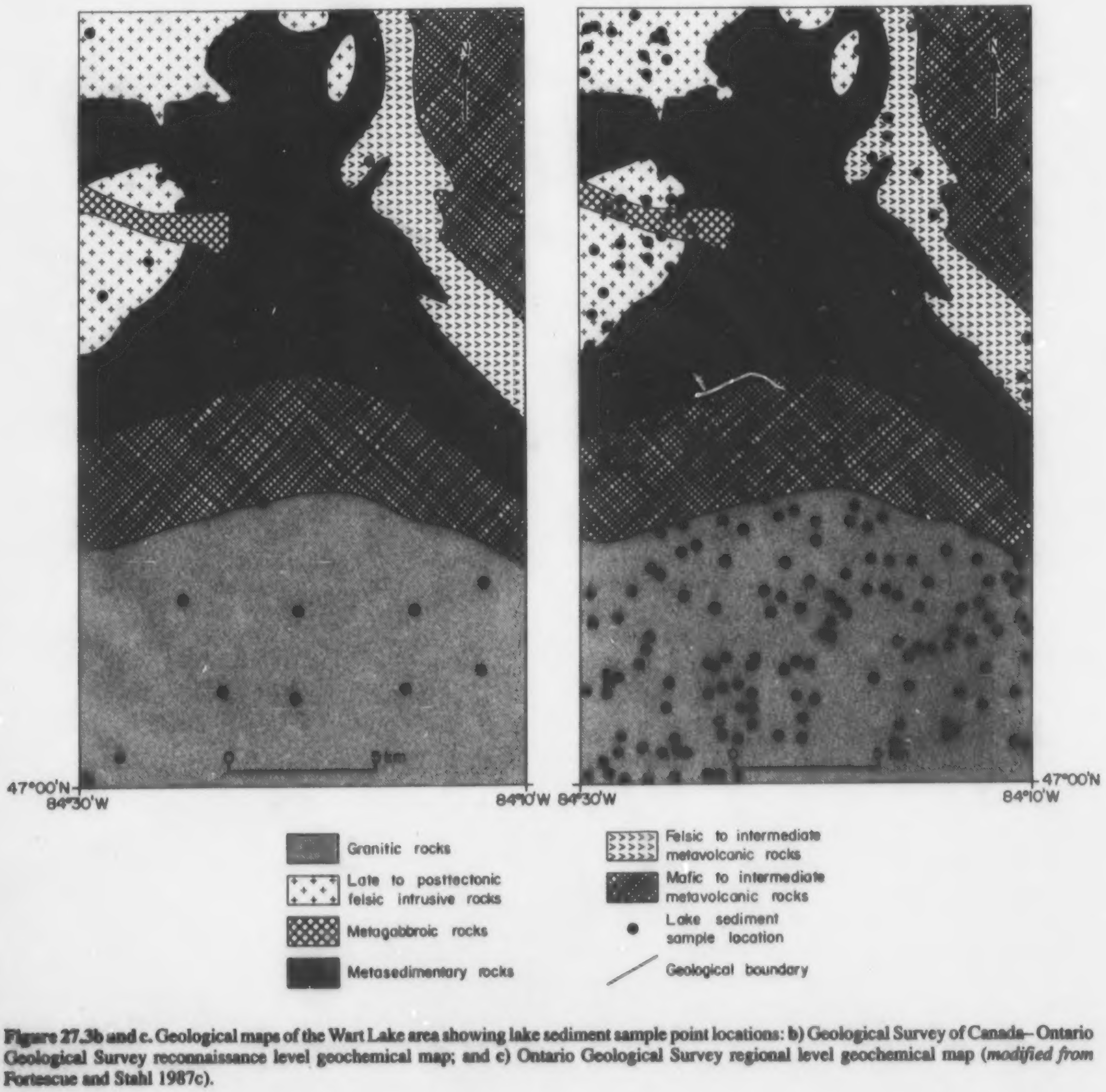

0
5 km
47°00'N
84°30'W
84°10'W
84°30'W
47°00'N
84°10'W
Granitic rocks
Late to posttectonic felsic intrusive rocks
Metagabbroic rocks
Metasedimentary rocks
Felsic to intermediate metavolcanic rocks
Mafic to intermediate metavolcanic rocks
Lake sediment sample location
Geological boundary

**Figure 27.3b and c.** Geological maps of the Wart Lake area showing lake sediment sample point locations: b) Geological Survey of Canada–Ontario Geological Survey reconnaissance level geochemical map; and c) Ontario Geological Survey regional level geochemical map (*modified from* Fortescue and Stahl 1987c).

geochemical mapping based on lake sediments and lakewaters is generally effective in landscapes with Precambrian bedrock and Quaternary cover shown as a, b, d and e on Figure 27.4. In contrast, geochemical mapping based on lake sediment is considered ineffective in areas where Precambrian bedrock is covered by a continuous layer of: 1) deep till (Figure 27.4c); 2) a complex overburden section (Figure 27.4f); 3) peat (Figure 27.4g); or 4) varved clay deposits (Figures 27.4h and 27.4i).

## Basal Till

The use of basal till as a sample medium for regional geochemical mapping was investigated by the OGS between 1978 and 1982, during the Kirkland Lake Initiatives Program (KLIP) (Fortescue, Lourim et al. 1984; Fortescue and Gleeson 1984). In the KLIP map area, a continuous cover of deep, complex and varied overburden (*see* Figures 27.4c, 27.4e, 27.4f, 27.4g, 27.4h and 27.4i) overlies Precambrian bedrock. Samples of basal till were obtained from either pits dug by backhoe or by reverse circulation overburden drilling.

## Stream Sediment

Stream sediment was the second sample medium tested by the OGS during the 1980s. In this case, regional geochemical mapping was carried out in drift-covered Paleozoic carbonate terrains of southwestern Ontario (Fortescue 1983a, 1984a).

## Lake Sediment

Modern geochemical mapping is designed to satisfy mineral resource appraisal and/or the need for base line information in enviromental geochemistry. An advantage of the OGS lake sediment regional geochemical mapping methodology (and the overbank low density geochemical surveys) is the simultaneous collection of pre- and post-anthropogenic material from each sample site.

The lake sediment regional geochemical mapping technique, developed by the OGS, is based on chemical data obtained from pre-*Ambrosia* lake sediment core material. This sample medium is known to have been deposited over 100 years ago (Dickman and Fortescue 1984; Fortescue, Dickman et al. 1984; Fortescue 1984b, 1985b, 1988a). The same lake sediment core can also be used to collect post-*Ambrosia* material, which accumulated at the bottom of the lake for 5 or more decades, prior to sample collection.

For example, in 1986, in the vicinity of the Sturgeon River (some 50 km northeast (downwind) of the Sudbury smelters; Figure 27.5), 139 lake sediment cores were collected from a 10 by 75 km sampling strip (Fortescue and Stahl 1987a). Pre- and post-*Ambrosia* data for lead, nickel,

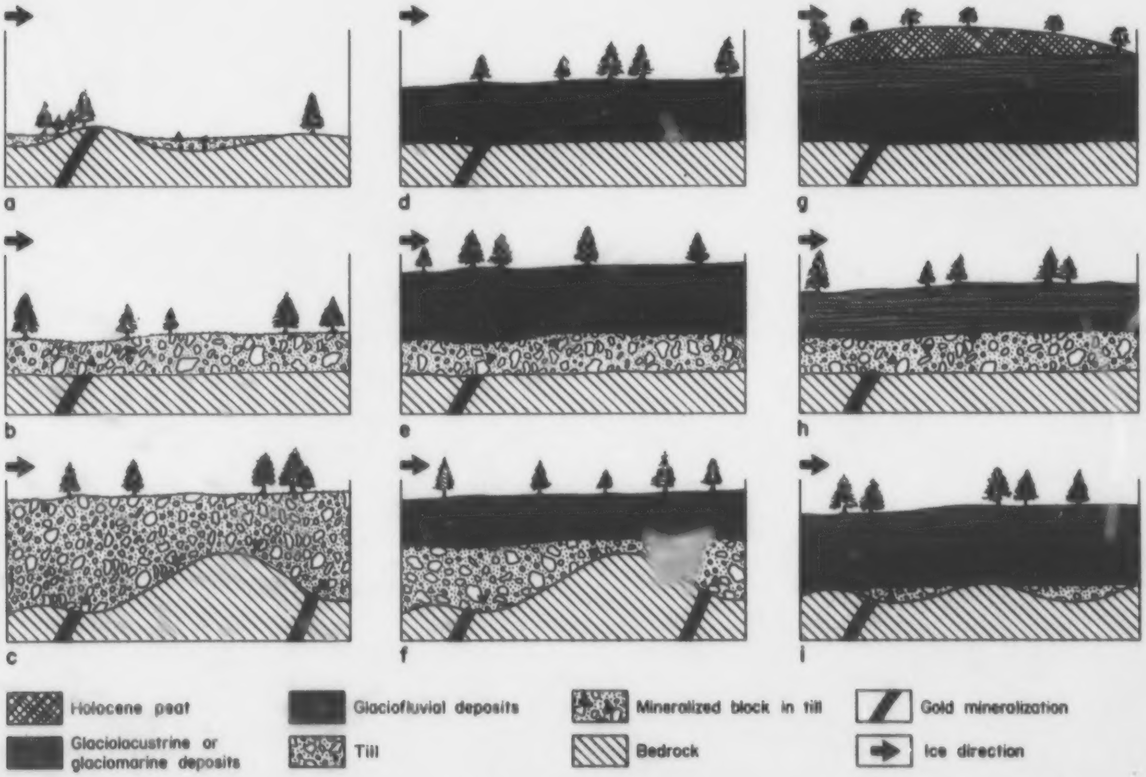


**Figure 27.4.** Cartoon sections of landscapes common in northern Ontario. The figure is designed as a guide to the choice of approaches to selection of sample media for regional geochemical mapping in the province (*modified from* Fortescue 1983c).

copper, strontium, lanthanum and uranium are plotted on Figure 27.5. These data show: 1) low levels for lead in the pre–*Ambrosia* material and enhanced levels in the post–*Ambrosia* material (due to atmospheric fallout from a non-point source; Figure 27.5a); 2) low levels of nickel and copper in the pre–*Ambrosia* material and enhanced levels in the post–*Ambrosia* material along a geochemical gradient away from the point source of atmospheric fallout in Sudbury (Figures 27.5b and 27.5c); 3) similar levels and patterns for strontium in pre– and post–*Ambrosia* material (Figure 27.5d); and 4) similar patterns for uranium in the pre– and post–*Ambrosia* material (Figure 27.5f). Figures 27.5d and 27.5e show patterns for strontium and lanthanum which are control elements not present at significant levels in atmospheric fallout.

Today, there is a need for the development of further regional geochemical mapping methodologies for use in areas of high mineral potential. These areas have: 1) continuous peat cover (*see* Figure 27.4g); 2) marl-rich lakes; and 3) complex Quaternary cover, which require a multimedia sampling approach (*see* Figures 27.4c to 27.4i).

## Summary

Sample point location and age of sample material are of paramount importance in geochemical mapping. The choice of sample medium is restricted by the landscape type in the area to be mapped. For example, in parts of Ontario where lakes and ponds are very common, lake sediment is the appropriate sample medium. In other areas with many streams and few lakes, stream sediment is the sample medium chosen for geochemical mapping. In either case, it is important that the medium is available over the entire map area and that the sample coverage is uniform.

Recently, the age of samples has become of considerable importance from the viewpoint of environmental geochemistry. The example from the Sturgeon River area showed how geochemical data obtained from dated lake sediments may provide important information for both mineral resource appraisal and environmental geochemistry.

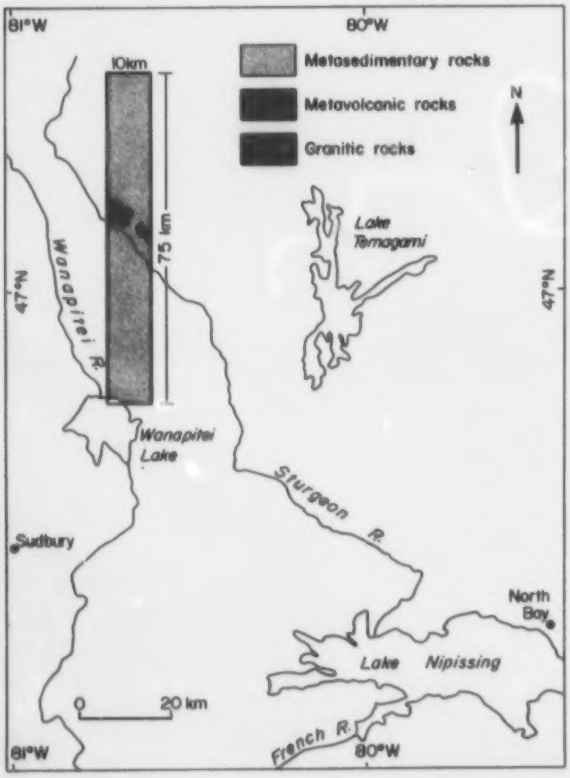


**Figure 27.5.** Location map shows geological setting of the Sturgeon Lake sampling strip.

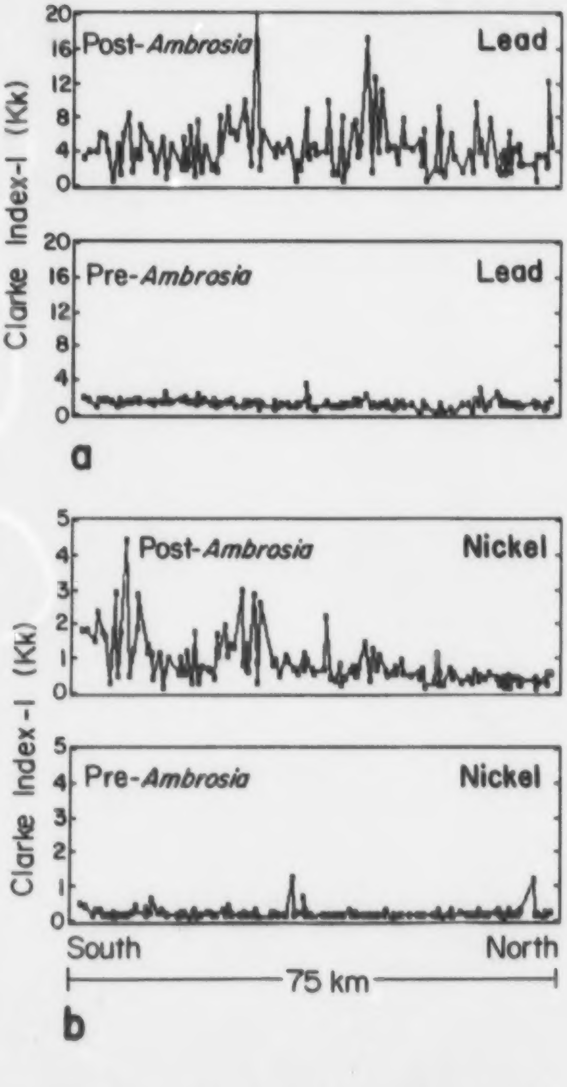


**Figure 27.5a and b.** Clarke-normalized element concentrations for post– and pre–*Ambrosia* data for 2 elements along the Sturgeon River sampling strip (*modified from* Fortescue and Stahl 1987a).

## RULE 3: STANDARDIZED METHODOLOGY

Experience in geochemical mapping in several countries has demonstrated that strict standardization of methods is essential for the preparation of reliable geochemical maps (Fortescue 1990b). Standardized methods used during the preparation of the Hanes Lake regional geochemical map are indicated on the flow diagram in Figure 27.6 (Fortescue and Vida 1990b).

During the planning stage (Figure 27.6a), lake sediment sample sites are positioned (and pre-numbered) on a 1:50 000 scale topographic map of the area to be mapped (Figure 27.6b). The number sequence includes randomly spaced missing numbers used for quality control purposes later on. Sample collection is carried out by 2 OGS staff working from the float of a helicopter (Figure 27.6c). They collect pre– (and post–) *Ambrosia* lake sediment core samples (Figure 27.6e) using the method described by Fortescue (1988b). A 1 L water sample is collected at all sample points (Figure 27.6d). At each sample site, the crew also note water colour, water depth, Secchi depth and other pertinent observations which may affect the analyses.

Using this method, 1 crew can sample 50 to 75 lake sites per day. At the base camp: 1) the pH of cooled water samples is measured within 24 hours of their collection; and 2) the sample bags are checked, sequenced and reference standards are inserted bearing the missing numbers (Figure 27.6f). At the end of a sampling program, the

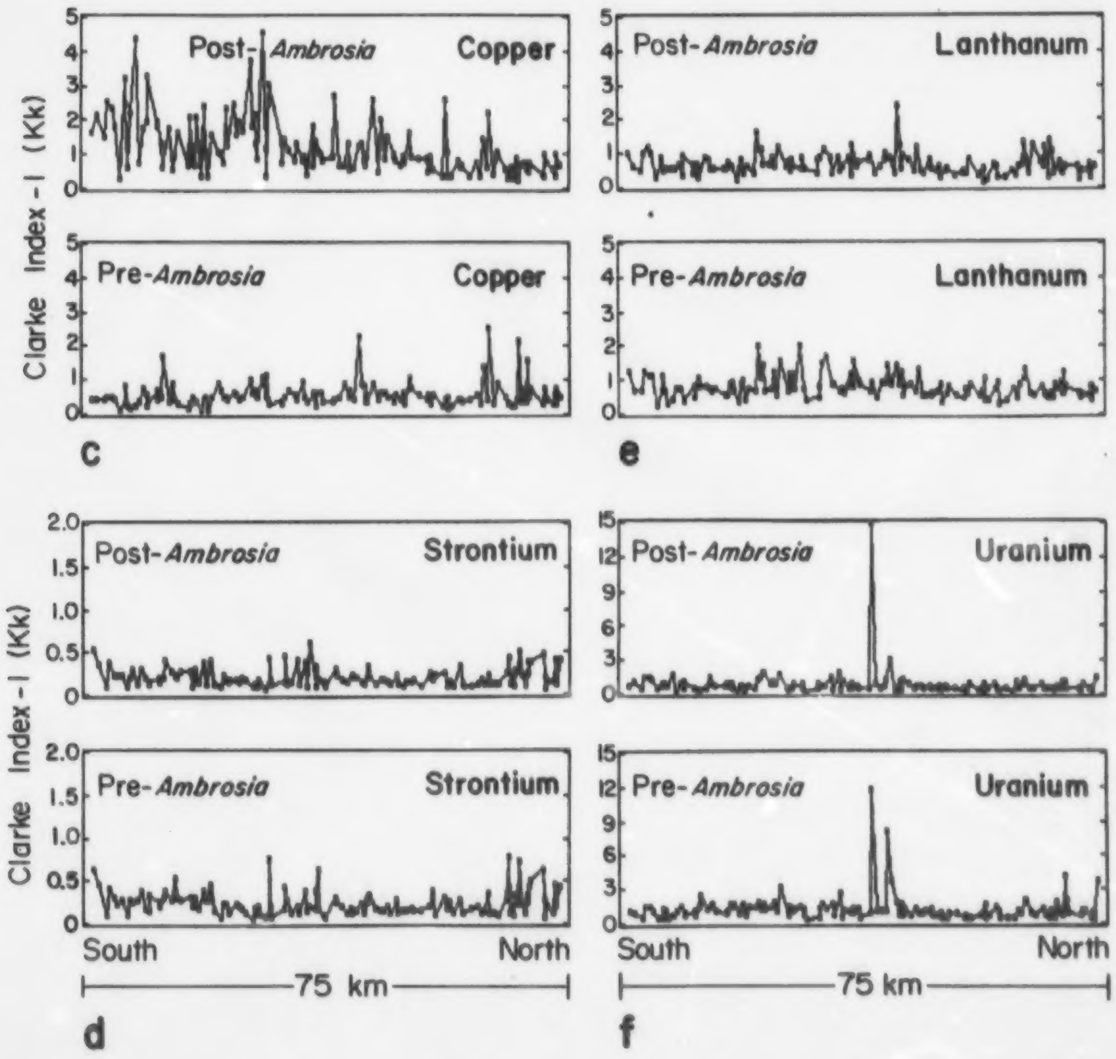
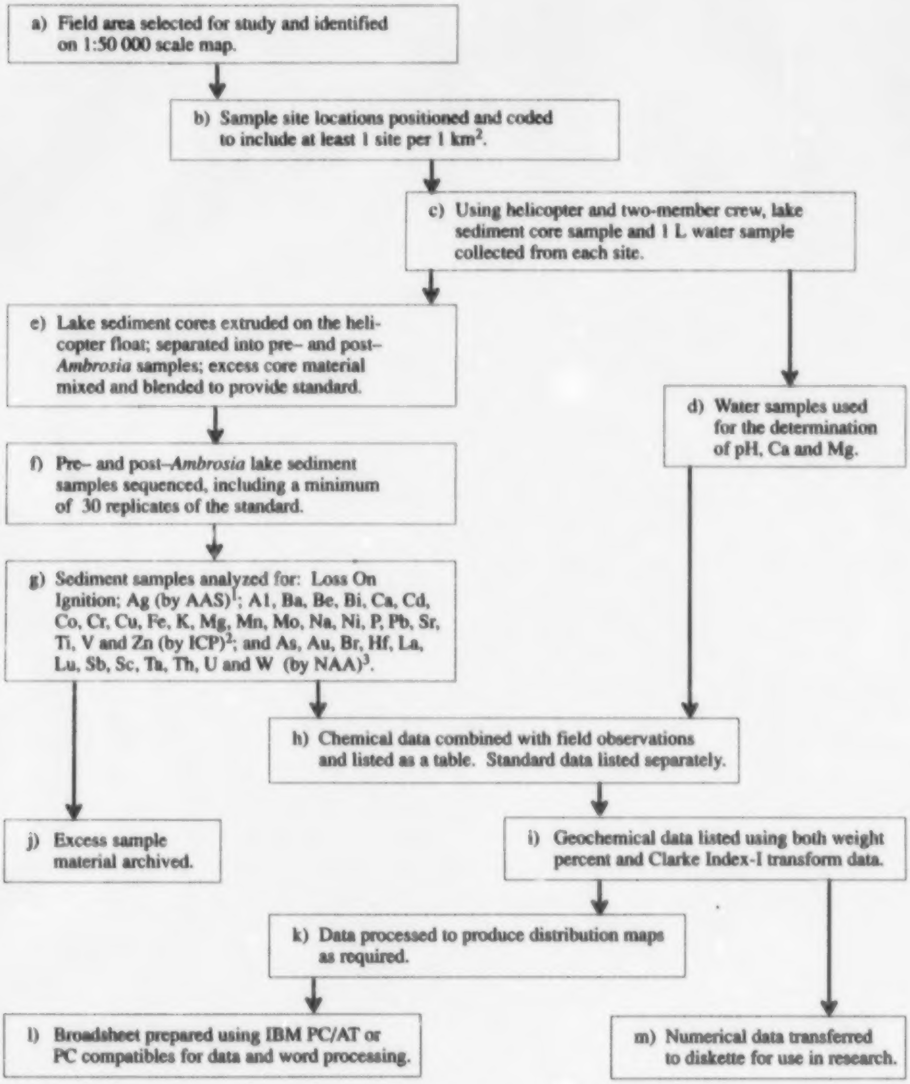


**Figure 27.5c to f.** Clarke-normalized element concentrations for post– and pre–*Ambrosia* data for 4 elements along the Sturgeon River sampling strip (*modified from* Fortescue and Stahl 1987a).



a) Field area selected for study and identified on 1:50 000 scale map.

b) Sample site locations positioned and coded to include at least 1 site per 1 km$^2$.

c) Using helicopter and two-member crew, lake sediment core sample and 1 L water sample collected from each site.

d) Water samples used for the determination of pH, Ca and Mg.

e) Lake sediment cores extruded on the helicopter float; separated into pre– and post–*Ambrosia* samples; excess core material mixed and blended to provide standard.

f) Pre– and post–*Ambrosia* lake sediment samples sequenced, including a minimum of 30 replicates of the standard.

g) Sediment samples analyzed for: Loss On Ignition; Ag (by AAS)[1]; Al, Ba, Be, Bi, Ca, Cd, Co, Cr, Cu, Fe, K, Mg, Mn, Mo, Na, Ni, P, Pb, Sr, Ti, V and Zn (by ICP)[2]; and As, Au, Br, Hf, La, Lu, Sb, Sc, Ta, Th, U and W (by NAA)[3].

h) Chemical data combined with field observations and listed as a table. Standard data listed separately.

i) Geochemical data listed using both weight percent and Clarke Index-I transform data.

j) Excess sample material archived.

k) Data processed to produce distribution maps as required.

l) Broadsheet prepared using IBM PC/AT or PC compatibles for data and word processing.

m) Numerical data transferred to diskette for use in research.

[1] Atomic Absorption Spectroscopy
[2] Inductively Coupled Plasma Mass Spectrometry
[3] Neutron Activation Analysis

**Figure 27.6.** Flow diagram of the methodology used for Ontario Geological Survey regional geochemical mapping based on lake sediment and water geochemistry (*modified from* Fortescue and Vida 1990b).

**Table 27.2.** Summary of quality control data obtained from a reference standard replicated at random 44 times within a batch of Trout Lake pre–*Ambrosia* lake sediment samples.

| Element | n | Mean (ppm) | Mean (Kk) | Minimum (Kk) | Maximum (Kk) | Range (Kk) | C.V.** (%) |
|---|---|---|---|---|---|---|---|
| LOI* | 44 | 46.25 | 46.25 | 44.20 | 47.60 | 3.40 | 1.56 |
| Sc | 44 | 9.56 | 0.38 | 0.34 | 0.42 | 0.07 | 4.08 |
| V | 44 | 21.07 | 0.15 | 0.14 | 0.17 | 0.03 | 5.06 |
| Al | 44 | 1.45 | 0.17 | 0.14 | 0.20 | 0.06 | 6.84 |
| Ti | 44 | 0.08 | 0.12 | 0.09 | 0.14 | 0.05 | 7.10 |
| Zn | 44 | 75.75 | 1.00 | 0.79 | 1.18 | 0.39 | 7.44 |
| Mn | 44 | 152.07 | 0.14 | 0.12 | 0.17 | 0.05 | 7.58 |
| Sr | 44 | 71.59 | 0.19 | 0.16 | 0.23 | 0.07 | 7.75 |
| Ba | 44 | 132.05 | 0.34 | 0.28 | 0.41 | 0.13 | 7.76 |
| P | 44 | 371.14 | 0.33 | 0.27 | 0.40 | 0.13 | 8.30 |
| Fe | 44 | 0.85 | 0.14 | 0.11 | 0.16 | 0.05 | 8.31 |
| K | 44 | 0.31 | 0.17 | 0.13 | 0.20 | 0.07 | 8.34 |
| La | 44 | 103.02 | 2.98 | 2.46 | 3.79 | 1.33 | 8.72 |
| Cu | 44 | 56.59 | 0.83 | 0.66 | 1.10 | 0.44 | 8.98 |
| Mg | 44 | 0.21 | 0.08 | 0.06 | 0.10 | 0.04 | 9.45 |
| Na | 44 | 0.33 | 0.14 | 0.11 | 0.18 | 0.06 | 10.19 |
| Co | 44 | 5.89 | 0.20 | 0.17 | 0.24 | 0.07 | 11.12 |
| Ca | 44 | 0.79 | 0.17 | 0.14 | 0.25 | 0.11 | 11.22 |
| U | 44 | 8.60 | 3.74 | 2.48 | 4.83 | 2.35 | 11.65 |
| Lu | 44 | 0.39 | 0.73 | 0.19 | 1.11 | 0.93 | 20.81 |
| Br | 44 | 5.80 | 2.32 | 1.60 | 4.32 | 2.72 | 21.58 |
| Cd | 44 | 1.47 | 9.16 | 6.25 | 15.63 | 9.38 | 22.55 |
| Th | 44 | 13.28 | 1.64 | 0.74 | 2.41 | 1.67 | 24.54 |
| Pb | 44 | 17.82 | 1.37 | 0.62 | 3.85 | 3.23 | 33.18 |
| Au | 44 | 4.75 | 1.19 | 0.25 | 2.25 | 2.00 | 34.62 |
| Mo | 44 | 1.52 | 1.27 | 0.83 | 3.33 | 2.50 | 41.26 |
| Ag | 44 | 0.68 | 8.52 | 6.25 | 18.75 | 12.50 | 42.11 |
| As | 44 | 1.77 | 0.98 | 0.56 | 2.22 | 1.67 | 45.30 |
| Ni | 44 | 33.75 | 0.34 | 0.20 | 1.33 | 1.13 | 51.75 |
| Cr | 44 | 41.48 | 0.34 | 0.15 | 1.38 | 1.23 | 63.20 |
| Sb | 44 | 0.25 | 1.24 | 0.50 | 5.00 | 4.50 | 85.07 |
| W | 44 | 4.05 | 3.37 | 1.67 | 21.67 | 20.00 | 121.56 |
| Hf | 44 | 3.75 | 1.34 | 0.71 | 16.79 | 16.07 | 179.96 |
| Be | 44 | 0.50 | 0.25 | 0.25 | 0.25 | 0.00 | 0.00 |
| Bi | 44 | 2.00 | 243.90 | 243.90 | 243.90 | 0.00 | 0.00 |
| Ta | 44 | 2.00 | 1.18 | 1.18 | 1.18 | 0.00 | 0.00 |

n Number of samples

C.V. Coefficient of Variation

* The LOI data are listed in percent (%).

** Data for elements with a coefficient of variation below 15% are normally accepted for advanced statistical analysis, image processing or insertion into a geographical information system without question. Data for elements with a coefficient of variation greater than 15% should be carefully inspected prior to interpretation by these methods.

sequenced bags of wet sediment are delivered to a contractor.

The contractor's laboratory first dries and ring pulverizes the sediment prior to chemical analysis. Then Loss On Ignition (LOI) and concentrations of 35 elements are determined in parts of each sample (Figure 27.6g). Data from chemical analyses are transferred directly from the contractor's laboratories to the OGS via a modem (Figure 27.6h). Calcium and magnesium levels in water samples are determined in the OGS Geoscience Laboratories (Figure 27.6d).

During the winter, field and laboratory data are combined into a single data base (Figure 27.6h). The chemical data for the quality control samples are then listed separately and analyzed statistically to determine the performance of the analytical methods (*see* Figure 27.6h). Data for an element are deemed acceptable for direct inputting to a GIS if the coefficient of variation (C.V.) of the quality control data is less than 16%. When the C.V. of an element in the quality control data is greater than 16%, caution is recommended if the data are to be included in a GIS.

As an example, data obtained from 44 replicates of a lake sediment standard sequenced randomly in a batch of 742 unknowns, are listed in Table 27.2. To facilitate inspection, data in Table 27.2 for 35 elements (and LOI) are ranked in order of increasing C.V. Table 27.2 shows that results for

**Table 27.3.** Clarke Index–I values (*modified from* Ranov and Yaroshevski 1972; Bowen 1979).

| Element | Clarke (ppm) | Element | Clarke (ppm) |
|---|---|---|---|
| Oxygen | 456000.0 | Gadolinium | 6.1 |
| Silicon | 273000.0 | Dysprosium | 5.0 |
| Aluminum | 83600.0 | Erbium | 3.5 |
| Iron | 62200.0 | Ytterbium | 3.1 |
| Calcium | 46600.0 | Hafnium | 2.8 |
| Magnesium | 27640.0 | Cesium | 2.6 |
| Sodium | 22700.0 | Bromine | 2.5 |
| Potassium | 18400.0 | Uranium | 2.3 |
| Titanium | 6320.0 | Europium | 2.1 |
| Phosphorous | 1120.0 | Tin | 2.1 |
| Manganese | 1060.0 | Beryllium | 2.0 |
| Fluorine | 544.0 | Arsenic | 1.8 |
| Barium | 390.0 | Tantalum | 1.7 |
| Strontium | 384.0 | Germanium | 1.5 |
| Sulphur | 340.0 | Holmium | 1.3 |
| Carbon | 180.0 | Molybdenum | 1.2 |
| Zirconium | 162.0 | Tungsten | 1.2 |
| Vanadium | 136.0 | Terbium | 1.2 |
| Chlorine | 126.0 | Thallium | 0.72 |
| Chromium | 122.0 | Lutecium | 0.54 |
| Nickel | 99.0 | Thulium | 0.50 |
| Rubidium | 78.0 | Iodine | 0.46 |
| Zinc | 76.0 | Indium | 0.24 |
| Copper | 68.0 | Antimony | 0.20 |
| Cerium | 66.0 | Cadmium | 0.16 |
| Neodymium | 40.0 | Mercury | 0.09 |
| Lanthanum | 35.0 | Silver | 0.08 |
| Yttrium | 31.0 | Selenium | 0.05 |
| Cobalt | 29.0 | Palladium | 0.02 |
| Scandium | 25.0 | Bismuth* | 0.0082 |
| Niobium | 20.0 | Tellurium | 0.0040 |
| Nitrogen | 19.0 | Gold | 0.0040 |
| Gallium | 19.0 | Ruthenium | 0.0010 |
| Lithium | 18.0 | Rhenium | 0.0007 |
| Lead | 13.0 | Platinum | 0.0005 |
| Praseodynium | 9.1 | Rhodium | 0.0002 |
| Boron | 9.0 | Osmium | 0.00002 |
| Thorium | 8.1 | Iridium | 0.000002 |
| Samarium | 7.0 | | |

* Caution: The value for bismuth (Bi) has been found in the author's experience to be set too low (Fortescue 1985a).

LOI and 18 elements (Al, Ba, Ca, Co, Cu, Fe, K, La, Mg, Mn, Na, P, Sc, Sr, Ti, U, V and Zn) obey the performance rule described above. Eleven of the 17 other elements (Ag, As, Au, Cd, Cr, Mo, Ni, Pb, Sb, Th and W) are considered acceptable only if caution is exercised during the interpretation of their map patterns. Data for 6 other elements (Be, Bi, Br, Hf, Lu and Ta) are considered unreliable for geochemical mapping purposes except where exceptionally high values (i.e., geochemical anomalies) are reported (*see* Table 27.2).

Let us now return to the description of Figure 27.6. Geochemical data included in OGS regional geochemical maps are standardized using 2 units, one chemical and the other geochemical. The chemical unit of measurement is weight percent (e.g., ppb, ppm and wt %). The geochemical unit is the Clarke Index-I (Figure 27.6i; Fortescue 1985a). The Clarke Index-I transform converts chemical data (obtained from the contractor; Figure 27.6h) into geochemical data. This is accomplished by dividing the weight % data for an element by an estimate of the abundance of that element in the earth's crust (Fortescue 1985a). The Clarke Index-I units (Kk), used by the OGS since 1980, are listed in parts per million (ppm) in Table 27.3. The use of Kk units simplifies the standardization of data tables, geochemical maps and diagrams included in hard copy releases of regional geochemical maps (Figure 27.6k). The use of Kk values also facilitates standardization of the marginal notes on these maps.

A data base on diskette(s) is available soon after the release of each regional geochemical map. The data base includes positional and geochemical data for the entire map in ASCII text or Lotus 1-2-3© format, suitable for immediate downloading into a GIS (Figure 27.6m).

## Types of Patterns on Geochemical Maps

In theory, geochemical maps describe all important element patterns resulting from mineral deposits or anthropogenic activity in the area mapped. Element patterns on geochemical maps are identified in 2 ways: 1) visually (by plotting, image processing and/or a GIS); and 2) mathematically (by using statistics).

Patterns on regional geochemical maps are either geochemical anomalies or geochemical provinces. On OGS regional geochemical maps, quotation marks around these terms indicate that the patterns were identified visually and the patterns have not been verified by resampling or reanalysis of the original samples.

## GEOCHEMICAL ANOMALY PATTERNS

Govette (1983, p.30) defined a geochemical anomaly in exploration geochemistry as:

> ...an abnormally high or low content of an element or element combination, or an abnormal spatial distribution of an element or element combination in a particular sample type in a particular environment as measured by a particular analytical technique.

On an OGS geochemical map, a geochemical anomaly is an isolated high value (or a small cluster of high values) where element level(s) is (are) usually at least 3 times greater than in the surrounding sites.

A excellent example of a geochemical anomaly occurs on the OGS Trout Lake regional geochemical map (Fortescue and Vida 1989). The Trout Lake map area is located in the central part of the Batchawana greenstone belt, northeast of Sault Ste. Marie (Figures 27.7 and 27.8). The geochemical anomaly occurs in 3 small lakes in Desbiens Township (Figure 27.7a). The anomaly pattern for zinc is shown on Figure 27.8b and that for copper on Figure 27.8c. Figures 27.8b and 27.8c are sawtooth plots of all zinc and copper values from 98 lake sites in a 5 km (west-east) map strip located along the eastern boundary of the Trout Lake map sheet.

## GEOCHEMICAL PROVINCES

Geochemical provinces are more common than anomaly patterns on geochemical maps. Levinson (1980, p.636) described geochemical province patterns as:

> ...large-scale crustal units characterized by common features of geological and geochemical evolution expressed in the chemical composition of the constituent geological complexes [formations], as well as in the endogenic and exogenic metalliferous and nonmetalliferous concentrations of the chemical elements.

On OGS regional geochemical maps, geochemical provinces are defined as areas of a geochemical map where the abundance of one (or more) elements is consistently higher, or lower, than in the surrounding area. In Ontario, geochemical provinces include linear and areal features which may involve one, or more elements. Also, areal geochemical provinces may include tens or hundreds of sample points.

Clarke Index-I transform data are standardized for plotting geochemical province and geochemical anomaly patterns on OGS regional geochemical maps. In the Trout Lake area, typical examples of geochemical provinces are found for uranium and arsenic. Mean Kk levels for uranium and arsenic in the entire Trout Lake area are 2.1 Kk (4.83 ppm) and 1.99 Kk (3.97 ppm), respectively. Geochemical provinces for uranium (features labelled A and B on Figure 27.9b)include all values greater than 10 Kk which are indicated by large dots. Large dots denoting arsenic values greater than or equal to 7.0 Kk describe similar patterns (features labelled A and B on Figure 27.9c). The arsenic map also includes a number of isolated geochemical anomalies (features labelled C, D, E, F and G on Figure 27.9c).

## SUMMARY

Two geochemical pattern types are common on OGS regional maps. These are geochemical provinces and geochemical anomalies. Marginal notes on OGS regional maps describe patterns considered to be important for mineral resource appraisal. Using the diskette provided with the hard copy map, other patterns in the 35 element geochemical data base may be examined by readers with access to a personal computer. The diskette may also be used for inputting the geochemical data into a GIS.

# PART III: APPROACHES TO REGIONAL GEOCHEMICAL MAPPING IN ONTARIO

## Using Basal Till as a Sample Medium

Lake or stream sediment regional geochemical mapping is not considered effective in areas covered with a continuous layer of deep and/or complex overburden (*see* Figures 27.4c, 27.4f to 27.4i). In these areas, the 3 rules of regional geochemical mapping previously described cannot be fully applied for 2 reasons.

1. It is impossible to sample the same geological formation across a region where a complex Quaternary section is combined with an uneven bedrock surface.
2. Drilling at a uniform sample density of 1 sample per km² is not practical because of the resources required.

A combination of these 2 constraints precludes the sampling of the same medium during regional geochemical mapping in areas of deep, continuous overburden.

Another problem is that deep overburden of the type found in landscapes c, f, g, h and i in Figure 27.4 is usually a product of mechanical (e.g., grinding) and physical weathering (e.g., frost wedging). Such material is usually unaffected by physical or chemical changes due to running water and/or the atmosphere. Physical weathering of this type almost always produces mixtures of locally and regionally derived material. This always complicates the interpretation of regional geochemical patterns based on material collected from the bedrock-overburden interface (e.g., basal till).

For example, a large percentage of locally derived material in a basal till weakens regional contributions and enhances local contributions to geochemical patterns and vice versa. Either way, from a regional viewpoint, the noise level of the geochemical data will be increased. Samples are subjected to a complex preparation procedure (summarized in Figure 27.10) in order to circumvent this problem and to obtain more detailed information from individual overburden samples. This procedure provides several geochemical and mineralogical data sets on which regional geochemical maps can be based.

In the late 1970s, when the Kirkland Lake Initiatives Program (KLIP) was commenced, it was believed that regional geochemical mapping would be feasible in the area. The increased volume of geochemical map data obtained from the basal till samples was thought to offset the irregular distribution (due to poor access) of drill-hole sites. At the end of the project, it was evident that this hypothesis was not tenable.

Sample preparation methodology (*see* Figure 27.10) was applied to 326 basal till samples (Fortescue and Gleeson 1984; Fortescue, Lourim et al. 1984). Each basal till sample produced: 1) 3 geochemical data sets obtained by analysis of different size fractions; 2) a gold grain count obtained by a visual examination; and 3) 4 sets of mineralogical data obtained from low power microscopic examinations of different subsamples (*see* Figure 27.10).

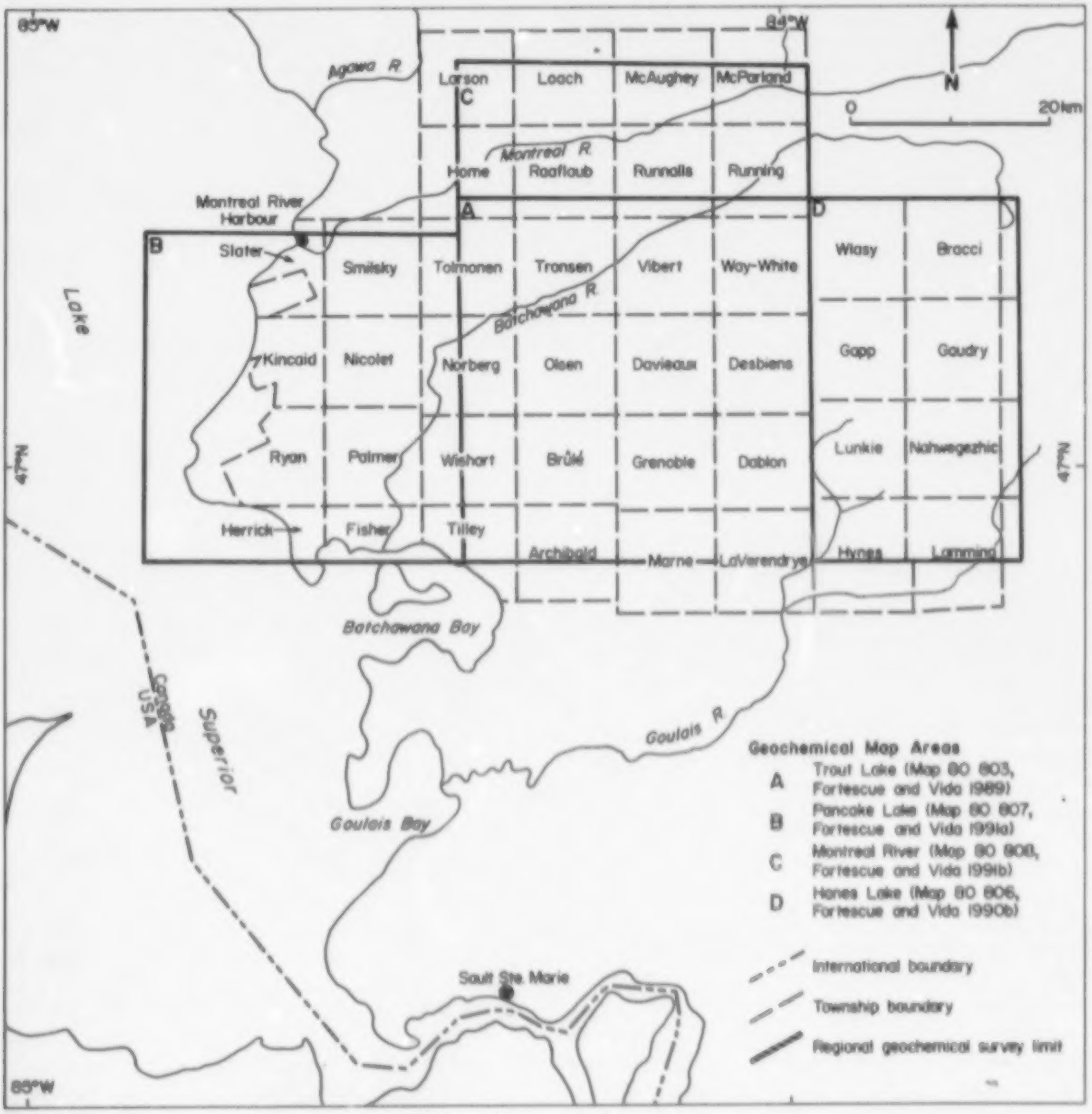

85°W
84°W
N
0
20 km
Agawa R.
Larson
Loach
McAughey
McParland
C
Montreal R.
Home
Roaflaub
Runnalls
Running
Montreal River Harbour
A
D
B
Slater
Smilsky
Tolmonen
Tronsen
Vibert
Way-White
Wlasy
Bracci
Batchawana R.
Lake
Kincaid
Nicolet
Norberg
Olsen
Davieaux
Desbiens
Gapp
Goudry
47°N
Ryan
Palmer
Wishart
Brûlé
Grenoble
Dablon
Lunkie
Nahwegezhic
47°N
Herrick
Fisher
Tilley
Archibald
Marne
LaVerendrye
Hynes
Lamming
Batchawana Bay
Canada
USA
Superior
Goulais R.
Geochemical Map Areas
A Trout Lake (Map 80 803, Fortescue and Vida 1989)
B Pancake Lake (Map 80 807, Fortescue and Vida 1991a)
C Montreal River (Map 80 808, Fortescue and Vida 1991b)
D Hanes Lake (Map 80 806, Fortescue and Vida 1990b)
Goulais Bay
Sault Ste. Marie
International boundary
Township boundary
Regional geochemical survey limit
85°W

**Figure 27.7a.** Index map showing the location of Ontario Geological Survey regional geochemical maps of the Batchawana greenstone belt 1989–1991 (*modified from* Fortescue and Vida 1991b).

For reasons explained above, it was extremely difficult to interpret most of the patterns on the KLIP regional geochemical and mineralogical maps. Also, in 1983, little was known of the details of the Quaternary stratigraphy in the KLIP area. In spite of this, some of the KLIP regional geochemical and mineralogical map patterns proved to be useful in mineral exploration.

For example, before the KLIP was completed, bedrock kimberlite had seldom been reported from the survey area (Lee 1968). However, as a result of the KLIP, kimberlite indicator minerals were found in many of the basal till samples.

The KLIP index map (Figure 27.11a) includes the limits of the area mapped, and the locations of drill holes and backhoe pits from which 326 basal till samples were collected (Fortescue, Lourim et al. 1984). During the project, a total of 243 geochemical and mineralogical maps of the area were prepared (Fortescue, Lourim et al. 1984). These included maps for kimberlite indicator minerals (spinel, chrome diopside and pyrope garnet; Figures 27.11b, 27.11c and 27.11d). A total of 120 spinels of 3 different colours were observed in basal till samples (*see* Figure 27.11b). Distribution maps for 69 black spinels, 26 pink spinels and 25 clear spinels showed no simple geochemical province pattern. Chrome diopside was found in 87 of the basal till samples. The provincial distribution for this mineral is similar to that for spinel, except that chrome diopside occurs more commonly to the southeast of the area (*see* Figures 27.11b and 27.11c). Unlike the patterns for spinel and chrome diopside, 83 of the 87 pyrope garnet occurrences are in the southern part of the map area (*see* Figure 27.11d).

In retrospect, the pyrope garnet pattern turned out to be of considerable importance. In the mid–1980s, 5 kimberlite pipes were found under deep overburden in Morrisette, Arnold and Clifford townships (*see* Figure 27.11d). If it is assumed that the sampled basal till was deposited by ice moving towards the southeast, the distribution of the pyrope

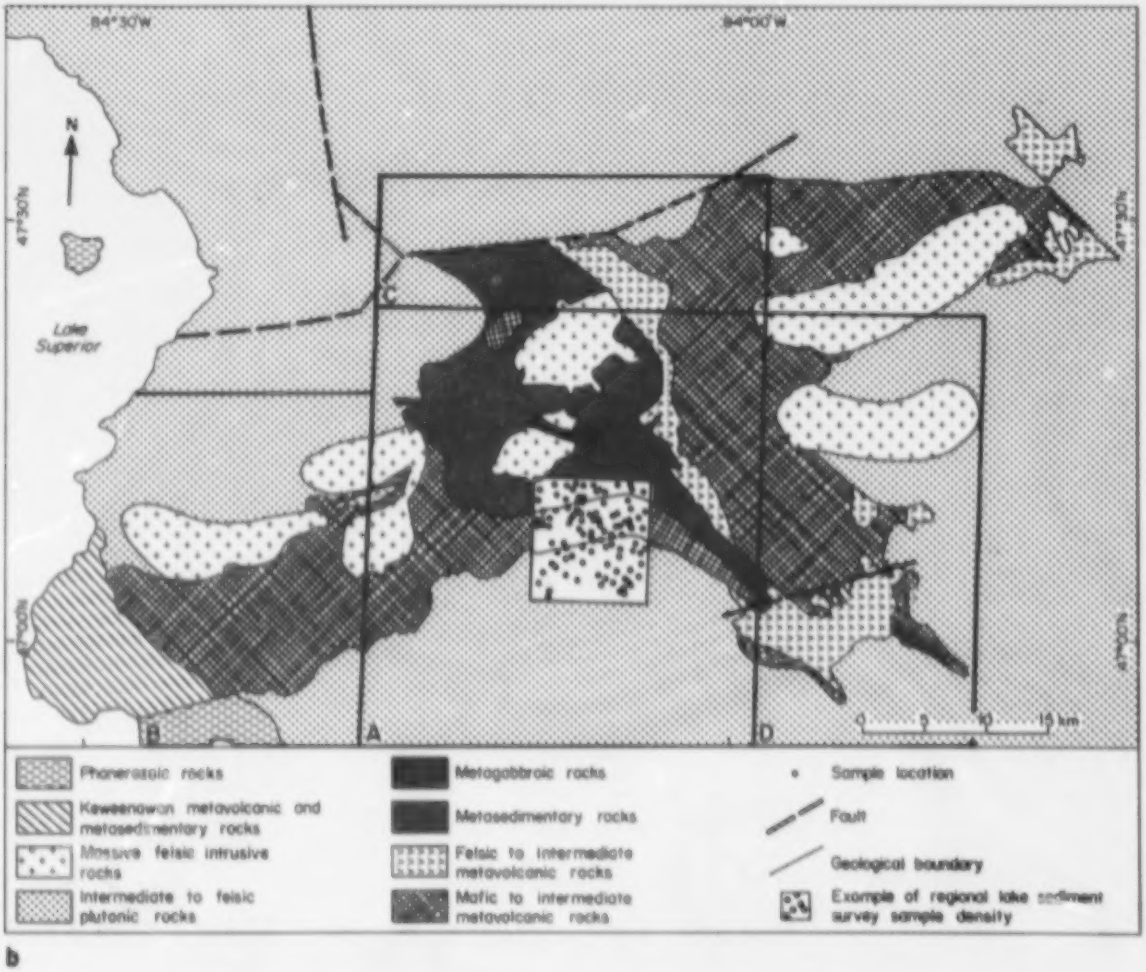

84°30'W
84°00'W
N
47°30'N
47°30'N
47°00'N
47°00'N
Lake Superior
C
B
A
D
0
5
10
15 km
Phanerozoic rocks
Keweenawan metavolcanic and metasedimentary rocks
Massive felsic intrusive rocks
Intermediate to felsic plutonic rocks
Metagabbroic rocks
Metasedimentary rocks
Felsic to intermediate metavolcanic rocks
Mafic to intermediate metavolcanic rocks
Sample location
Fault
Geological boundary
Example of regional lake sediment survey sample density
b

Figure 27.7b. Generalized geological map of the Batchawana greenstone belt. Note the density of sample locations in comparison to the size of the study area.

garnet can be related directly to the kimberlite pipes as all but 4 of the samples containing pyrope garnet were collected southeast of the pipes. Diamonds of various sizes have been reported from all of the kimberlite pipes (W.C. Fipke, Dia Met Minerals Ltd., personal communication, 1990).

### COMMENTS

Regional geochemical mapping using basal till samples obtained from boreholes now appears impractical due to: 1) complex Quaternary stratigraphy in many areas of deep overburden; 2) the problem of drill site access in remote areas; and 3) the expense of deep overburden sampling at a uniform density of 1 sample per km². Detailed geochemical mapping can be an important component of the OGS mineral resource appraisal process if overburden drilling is both carried out in areas of exceedingly high mineral potential and based on detailed Quaternary stratigraphy of the overburden sections. A recent study by Bajc (1988) clearly supports this assertion.

## Using Stream Sediments as a Sample Medium

In 1981, the OGS completed the Southwestern Ontario Regional Geochemical Survey in an area of 25 000 km² extending from the Niagara River to Lake Huron (Figure 27.12a). The project involved the collection of over 4000 stream sediment samples in which concentrations of 28 elements were determined.

The Southwestern Ontario Geochemical Survey area was chosen on the basis of geological considerations for a test of stream sediment regional geochemical mapping methodology. The area is underlain by Paleozoic sedimentary strata which strike northwest (Figure 27.12b). The last ice movement across the area was towards the southwest (Hewitt 1972). Quaternary deposits in the survey area are of 3 types: 1) tills and moraines in the western part of the area; 2) glaciofluvial sands and gravels in the centre of the area; and 3) lake sediments in the eastern part of the area (Figure 27.12c).

The Southwestern Ontario Geochemical Survey mapping project was designed to fulfill the requirements of the 3 regional geochemical mapping principles discussed above. The overall aim of the survey was to provide a regional geochemical data base which could be utilized for mineral resource appraisal and base line environmental geochemistry.

The first principle of geochemical mapping was satisfied in the Southwestern Ontario Geochemical Survey area as follows: 1) the entire area to be mapped was divided into 10 by 10 km squares (micromodules); 2) each micromodule was further divided into four 5 by 5 km squares (micromodule quarters); and 3) 5 stream sediment sample points were located within each micromodule quarter. Limits of micromodules, micromodule quarters and sample sites were plotted on 1:50 000 topographic base maps that were later used for field sampling. All stream sediment collection sites were positioned at least 50 m upstream from

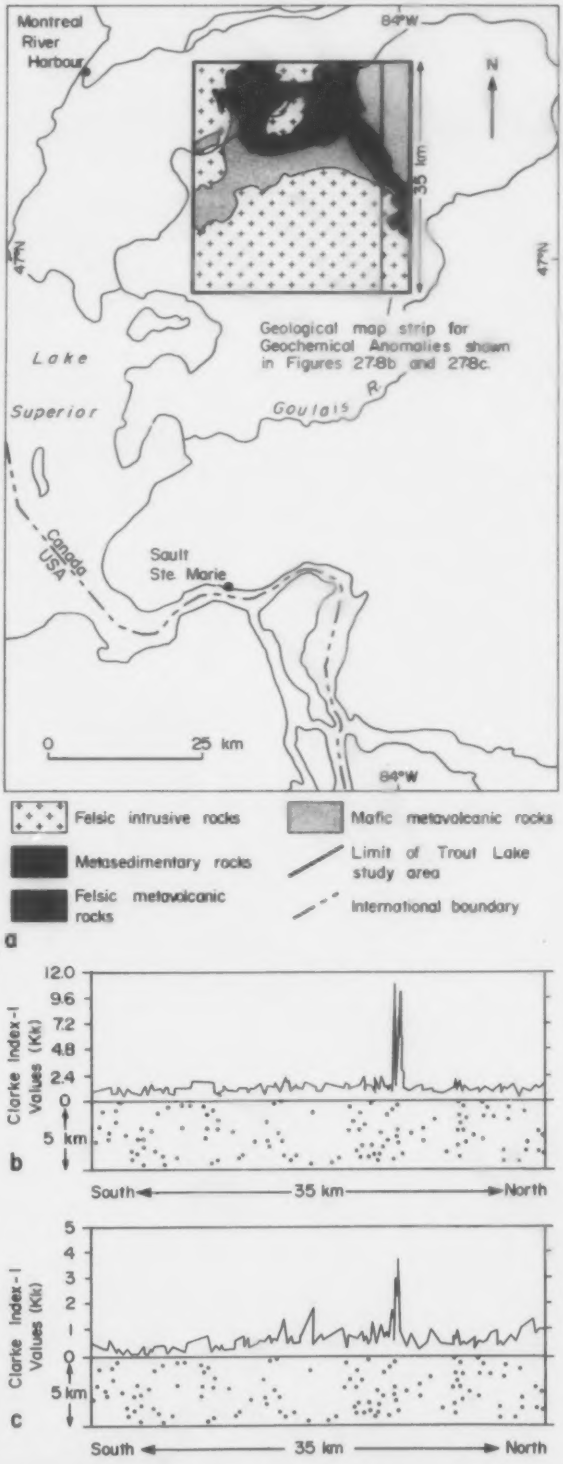
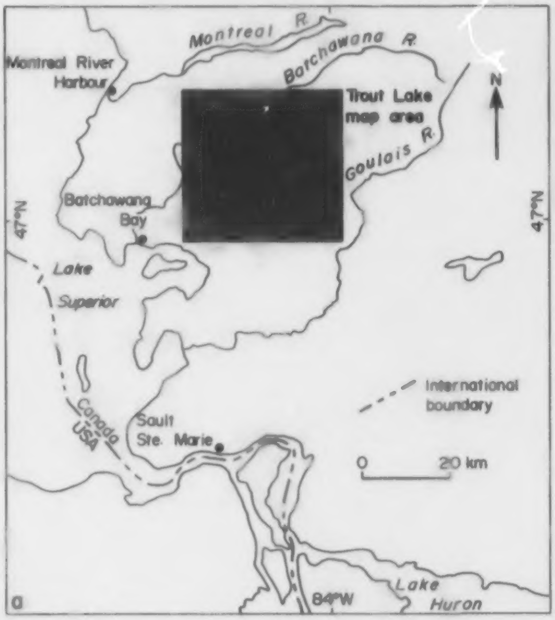

Montreal
River
Harbour
84°W
N
35 km
47°N
47°N
Geological map strip for Geochemical Anomalies shown in Figures 27.8b and 27.8c.
Lake
Superior
Goulais R.
Canada
USA
Sault
Ste. Marie
0
25 km
84°W
Felsic intrusive rocks
Mafic metavolcanic rocks
Metasedimentary rocks
Limit of Trout Lake study area
Felsic metavolcanic rocks
International boundary
a
12.0
9.6
7.2
4.8
2.4
0
Clarke Index-I Values (Kk)
5 km
b
South
35 km
North
5
4
3
2
1
0
Clarke Index-I Values (Kk)
5 km
c
South
35 km
North

**Figure 27.8.** a) Generalized map showing the location and geology of the Trout Lake study area. b) Geochemical section of a zinc anomaly in the Trout Lake area. c) Geochemical section of a copper anomaly in the Trout Lake area. For parts b and c, the bottom panel of each section represents the geological map strip located at the eastern edge of the Trout Lake study area. The dots represent sample locations (*modified from* Fortescue and Vida 1989).

road bridges. Prior to this mapping project, a sample density of 1 sample per 5 km² was considered adequate for regional geochemical mapping in southwestern Ontario.

Constraints of the second principle of regional geochemical mapping were met by collecting samples of active stream sediment. This material is found over the entire Southwestern Ontario Geochemical Survey area. When the survey was planned, it was expected that fine-blown eolian mineral matter, derived from arable fields adjacent to streams, might blur the geochemical map patterns. It was discovered that this blurring effect was insignificant during the preliminary interpretation of patterns in the geochemical data. Stream sediment samples were dried, crushed and passed through an 80 mesh (180 micron) sieve before chemical analysis.


Montreal R.
Montreal River Harbour
Batchawana R.
Trout Lake map area
N
Goulais R.
Batchawana Bay
47°N
47°N
Lake Superior
International boundary
Canada
USA
Sault Ste. Marie
0 20 km
84°W
Lake Huron
a

The third principle of regional geochemical mapping was obeyed by standardization of all the steps in the preparation of the Southwestern Ontario Geochemical Survey maps. Data for levels of 28 elements (including Al, Ag, Ba, Be, Ca, Co, Cr, Cu, Fe, K, Mg, Mn, Na, Ni, P, Pb, Sr, Th, Ti and Zn) were obtained from each of the 4000 samples collected during the survey. The project also included analyses of several hundred standard samples inserted within the unknowns for quality control purposes. At the time of writing, none of the geochemical patterns in the Southwestern Ontario Geochemical Survey data had been verified either by resampling or reanalysis of the original samples.

## THE SOUTHWESTERN ONTARIO GEOCHEMICAL SURVEY MAPS

Major geochemical anomaly and geochemical province patterns in the data sets for 20 selected elements were presented and described on 2 broadsheets (Fortescue 1983a, 1984a).

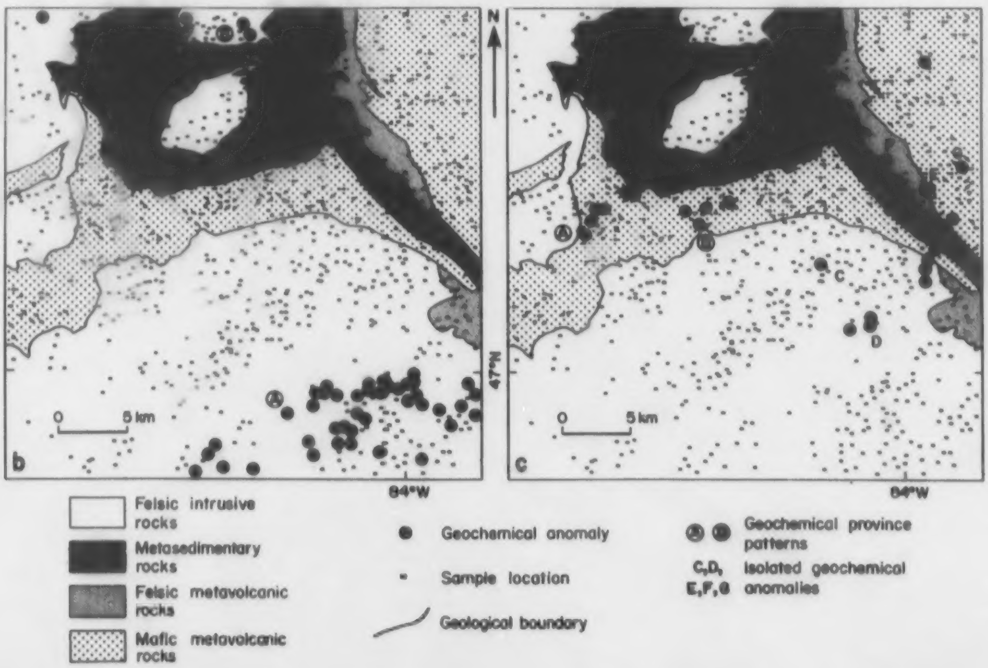
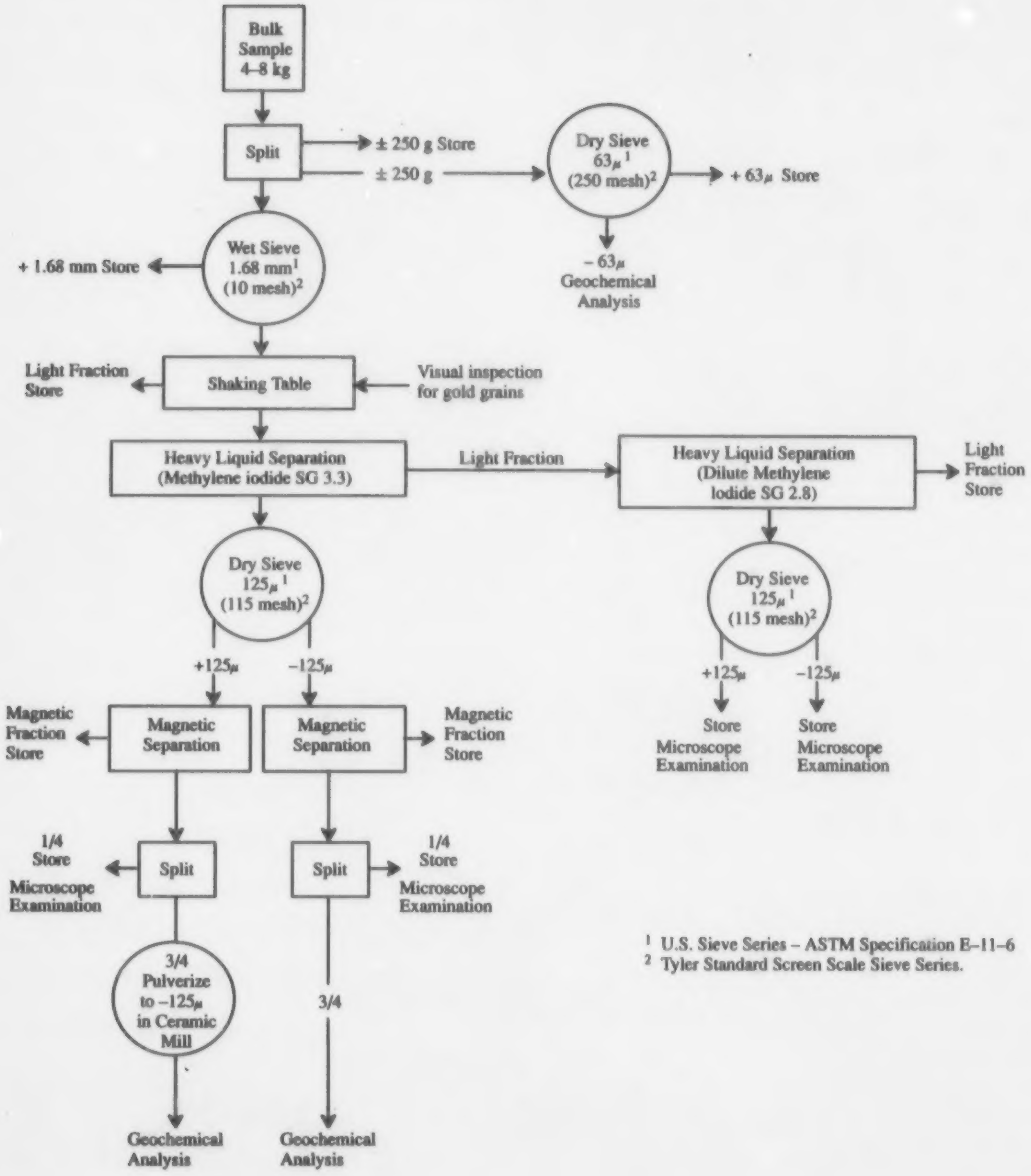

N
47°N
0 5 km
b
A
0 5 km
c
A
B
C
D
E
F
G
84°W
84°W
Felsic intrusive rocks
Metasedimentary rocks
Felsic metavolcanic rocks
Mafic metavolcanic rocks
Geochemical anomaly
Sample location
Geological boundary
Geochemical province patterns
C,D, E,F,G isolated geochemical anomalies

Figure 27.9. a) Generalized map showing the location of the Trout Lake study area. Geochemical province patterns for: b) uranium; and c) arsenic, as related to the general geology and sample locations in the Trout Lake study area (*modified from* Fortescue and Vida 1989). Geochemical provinces are denoted by Clarke Index-I values greater than or equal to 10 Kk for uranium and 7.0 Kk for arsenic.


Bulk Sample 4–8 kg
Split
± 250 g Store
± 250 g
Dry Sieve 63µ¹ (250 mesh)²
+ 63µ Store
– 63µ Geochemical Analysis
Wet Sieve 1.68 mm¹ (10 mesh)²
+ 1.68 mm Store
Light Fraction Store
Shaking Table
Visual inspection for gold grains
Heavy Liquid Separation (Methylene iodide SG 3.3)
Light Fraction
Heavy Liquid Separation (Dilute Methylene Iodide SG 2.8)
Light Fraction Store
Dry Sieve 125µ¹ (115 mesh)²
+125µ
–125µ
Store Microscope Examination
Store Microscope Examination
Magnetic Fraction Store
Magnetic Separation
Magnetic Separation
Magnetic Fraction Store
1/4 Store Microscope Examination
Split
Split
1/4 Store Microscope Examination
3/4 Pulverize to –125µ in Ceramic Mill
3/4
Geochemical Analysis
Geochemical Analysis

[1] U.S. Sieve Series – ASTM Specification E-11-6
[2] Tyler Standard Screen Scale Sieve Series.

**Figure 27.10.** Flow diagram of sample preparation for geochemical and mineralogical analysis of the Kirkland Lake Initiatives Program (KLIP) basal till samples (*modified from* Fortescue and Gleeson 1984).

Figure 27.12 includes maps of the Southwestern Ontario Geochemical Survey area showing: a) geography; b) bedrock geology; c) Quaternary geology; d) the micromodule grid (with points plotted at the centre of each micromodule); and e) the micromodule-quarter grid (with points plotted at the centre of each micromodule quarter).

The Southwestern Ontario Geochemical Survey geochemical anomaly and province patterns were displayed using micromodule (i.e., 20 sample) composite values on the 1983 broadsheet (Fortescue 1983a). Using the procedure described above, parts per million (ppm) values for each element were transformed into Clarke Index-I units (Kk). The resulting Kk value data sets for all 20 elements were classified using the same Kk frequency distribution cell limits. Values for an element in each cell of the frequency distribution were plotted as large dots on a copy of the micromodule grid map (Figure 27.12d). The resulting frequency distribution cell maps for an element were then arranged in a column with the lowest Kk value map at the bottom and the highest Kk value map at the top. The columns were then ranked in order of increasing median Kk value to produce a scan sheet (Figure 27.13). The ranking procedure facilitates visual comparisons among the element Kk level map patterns. Surprisingly, it is easy to identify many geochemical map patterns on Figure 27.13 such as: 1) geochemical gradient; 2) geochemical anomaly; and 3) geochemical province patterns. The reader should refer to Figure 27.13 for a description of these patterns.

The second Southwestern Ontario Geochemical Survey broadsheet (Fortescue 1984a) included regional geochemical map patterns similar to those just described, but based on micromodule-quarter composites (e.g., mean values for 5 samples). A scan sheet format was not used to display geochemical patterns in the 1984 broadsheet. Instead, regional geochemical patterns involving both micromodule and micromodule-quarter plots were combined on the same map (e.g., zinc in Figure 27.14).

### Geochemical Patterns

The median level for the 727 micromodule quarters in the Southwestern Ontario Geochemical Survey map area is around 1.25 Kk (100 ppm) Zn. Eleven of these samples were found to have Zn levels greater than 10 Kk (760 ppm) and an additional 5 samples were found to have levels between 5 and 9.9 Kk (380 and 759 ppm; *see* Figure 27.14). When all samples with Zn levels greater than or equal to 5 Kk are plotted, high values form a geochemical province pattern in the north-central part of the map area (*see* Figure 27.13, pattern 2; Figure 27.14). This province pattern for zinc covers an area of approximately 400 km² and is probably due to small amounts of zinc in the bedrock which have been concentrated locally in swampy areas.

An interesting geochemical anomaly pattern for lithium, which is so far unexplained, occurs in the Niagara Peninsula (Figure 27.15). The Southwestern Ontario Geochemical Survey median value for Li is between 0.75 and 1.0 Kk (13.50 and 18.00 ppm). The Li geochemical province includes 61 samples with levels greater than or equal to 2.50 Kk (45.00 ppm; *see* Figure 27.15). Three geochemical maps are used to describe the lithium geochemical province (*see* Figures 27.15a, 27.15b and 27.15c). They include data for lithium in micromodule and micromodule-quarter composites in the area. The accompanying table in Figures 27.15a, 27.15b and 27.15c shows the frequency distribution of lithium Clarke Index-I codes. It includes 6 sample points with Li values greater than or equal to 3 Kk (54 ppm). Four of these points appear as a weak linear anomaly in the western part of the area, whereas the other 2 points occur near the Buffalo–Niagara Falls industrial complex (*see* Figure 27.15a). Figures 27.15b and 27.15c include plots of Li levels greater than or equal to 2.75 Kk (49.5 ppm) and 2.50 Kk (45 ppm), respectively. Figure 27.15c also includes the pattern for all Li values greater than or equal to 2.0 Kk (36 ppm), because these are associated with the stronger Li map patterns just described.

In summary, a lithium geochemical province pattern was found in the stream sediment composite data collected during the Southwestern Ontario Geochemical Survey project in the Niagara Peninsula. This geochemical province is a weak northeast-trending linear anomaly. At the time of writing in early 1991, the pattern had not been explained. The pattern is coincident with aeromagnetic linear anomaly features (*see* Map 2587, map case).

Geochemical province patterns for potassium, aluminum and beryllium (Figures 27.16a, 27.16b and 27.16c) are associated with glacial lake sediment cover of the Niagara Peninsula (*see* Figure 27.12c). High levels of argillaceous minerals in the lake sediments probably account for geochemical province patterns for potassium and aluminum in the area (*see* Figures 27.16a and 27.16b). The beryllium geochemical province is probably due to this element having almost the same ionic potential as aluminum, causing the 2 elements to behave similarly during the formation of secondary phyllosilicates (Goldschmidt 1954).

## SUMMARY

Even though they are based on composite samples, the Southwestern Ontario Geochemical Survey maps show both geochemical anomaly and geochemical province patterns. Although the composite of the regional geochemical data was carried out for practical reasons, it was concluded that important geochemical map patterns can still be delineated using micromodules and micromodule-quarter composites. More important, the survey demonstrated that it is feasible to produce valid geochemical maps based on composite samples. If the samples are composited prior to chemical analysis, then geochemical maps can be made rapidly and cheaply, using very few chemical analyses.

## Using Lake Sediments and Waters as Sample Media

By 1991, 5 regional geochemical maps had been released by the OGS based on pre-*Ambrosia* lake sediment (Fortescue and Vida 1989, 1990a, 1990b, 1991a, 1991b). Four of these maps are located in the Batchawana greenstone belt (*see* Figure 27.7) and one in the Michipicoten greenstone belt (Figure 27.17).

Examples of zinc, copper, uranium and arsenic geochemical patterns from the Trout Lake map were described above in the section entitled Geochemical Patterns (Fortescue and Vida 1989). This section is concerned with other pattern types important for interpreting modern regional geochemical maps.

## LINEAR GEOCHEMICAL PROVINCES

Figures 27.18a through 27.18g present geochemical-geological map strips (G-GMS) displaying gold data for 946 Trout Lake area pre–*Ambrosia* samples. In Figure 27.18, the 5 km wide G-GMS are oriented from south to north (left to right). The western margin of the Trout Lake map area is included in Figure 27.18a and the eastern margin in Figure 27.18g. The mean level for gold in the entire Trout Lake area is 0.71 Kk (2.81 ppb). The gold linear geochemical province of interest is highlighted by arrows on

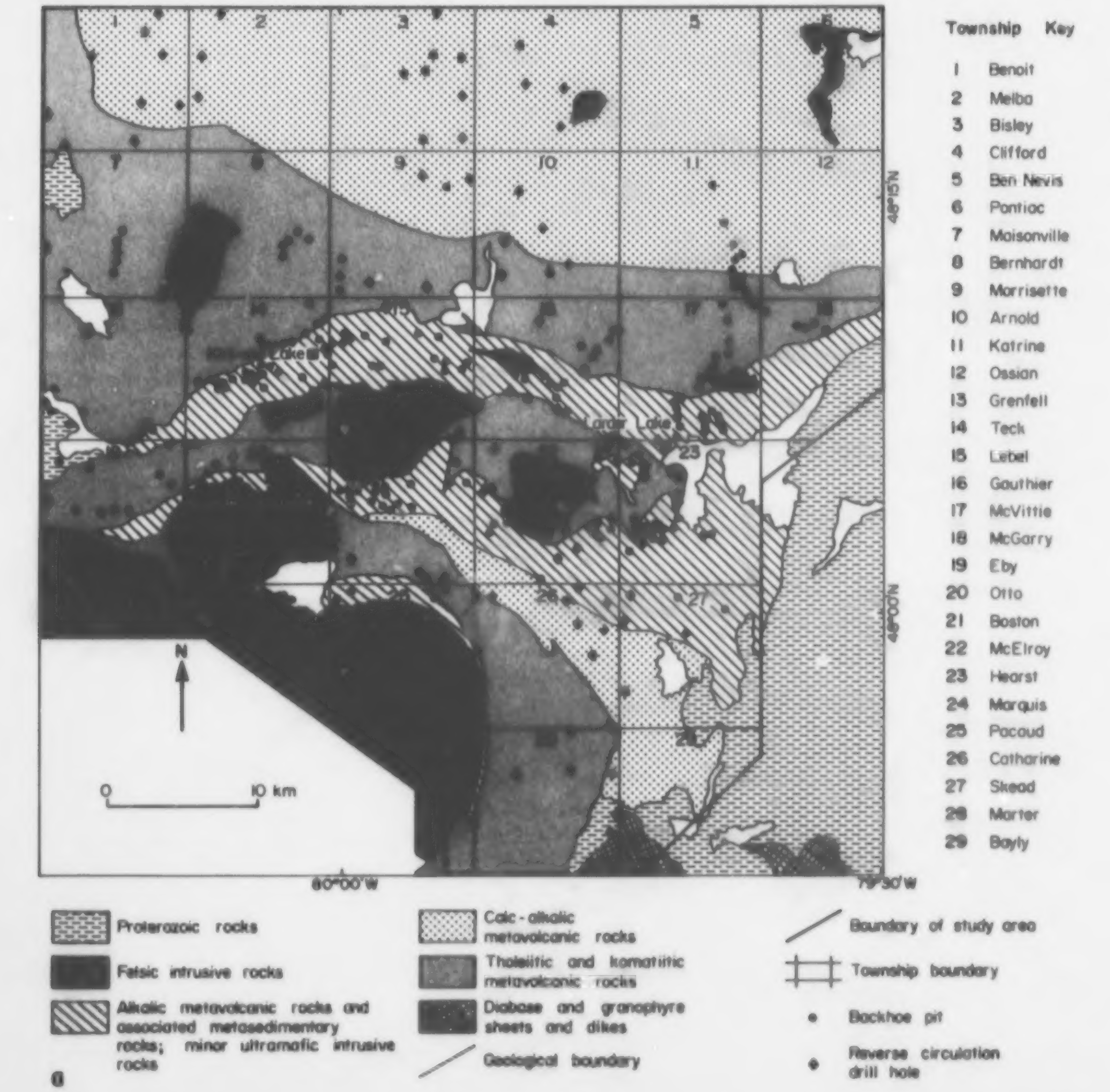
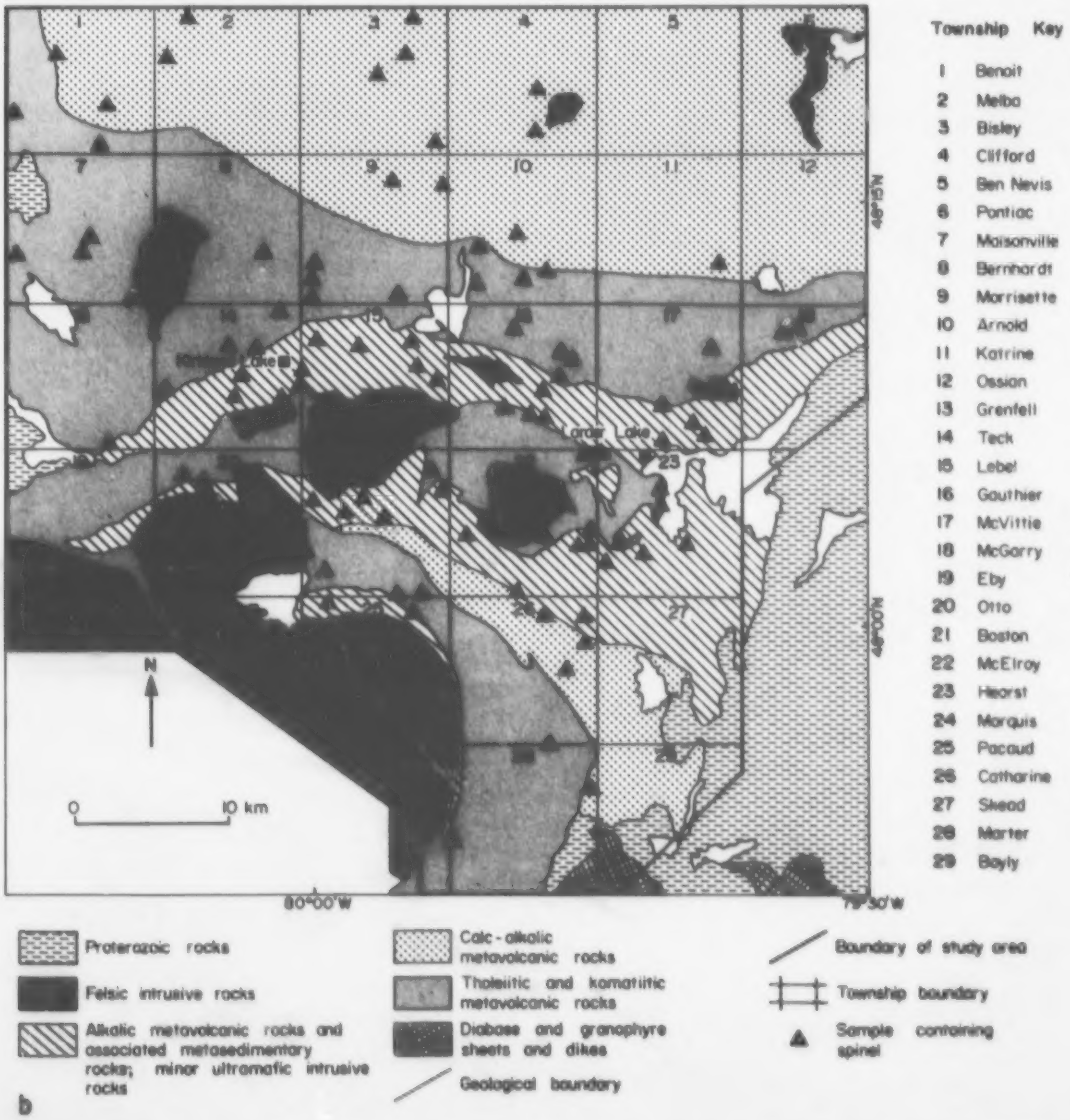

1
2
3
4
5
6
7
9
10
11
12
14
15
17
23
26
27
Kirkland Lake
Larder Lake
48°15'N
48°00'N
80°00'W
79°30'W
N
0
10 km
Township Key
1 Benoit
2 Melba
3 Bisley
4 Clifford
5 Ben Nevis
6 Pontiac
7 Maisonville
8 Bernhardt
9 Morrisette
10 Arnold
11 Katrine
12 Ossian
13 Grenfell
14 Teck
15 Lebel
16 Gauthier
17 McVittie
18 McGarry
19 Eby
20 Otto
21 Boston
22 McElroy
23 Hearst
24 Marquis
25 Pacaud
26 Catharine
27 Skead
28 Marter
29 Bayly
Proterozoic rocks
Felsic intrusive rocks
Alkalic metavolcanic rocks and associated metasedimentary rocks; minor ultramafic intrusive rocks
Calc-alkalic metavolcanic rocks
Tholeiitic and komatiitic metavolcanic rocks
Diabase and granophyre sheets and dikes
Geological boundary
Boundary of study area
Township boundary
Backhoe pit
Reverse circulation drill hole
a

**Figure 27.11a.** Geological maps of the Kirkland Lake Initiatives Program study area displaying general geology and basal till sample site locations (*modified from* Fortescue and Gleeson 1984).

Figures 27.18b–e. This linear feature parallels the strike of the greenstone belt for 20 km and includes 4 gold values ranging from 2 to 2.5 Kk (8 to 10 ppb). The gold pattern in lake sediment is associated with a small, but persistent, iron formation in the greenstone underlying lake catchment areas. This iron formation has been reported to have low gold values throughout its length (G. Bennett, Resident Geologist's office, Sault Ste. Marie, Mines and Minerals Division, personal communication, 1989). It is noteworthy that the OGS regional geochemical mapping methodology is sensitive enough to trace this relatively weak gold pattern over a distance of 20 km. Two single sample point geochemical anomalies for gold also occur in Figure 27.18. Anomaly A (64 ppb Au) is in Figure 27.18f; Anomaly B (21 ppb Au) is indicated in Figure 27.18c. At the time of writing, these results had not been verified by further analysis.

The gold linear geochemical province just described was concordant with the strike of the greenstone belt. This contrasts with a weak, 3 sample point discordant thorium linear geochemical province, which was also found in the Trout Lake map data (Figures 27.19e, 27.19f and 27.19g). In this case, Th values of 2.72 Kk (18 ppm), 1.73 Kk (14 ppm) and 1.52 Kk (12 ppm) form the 15 km long linear province where the background level for Th is 0.39 Kk (3.2 ppm). This thorium geochemical linear province is assumed to be due to a structural feature which cuts across the regional strike of the greenstone belt.



Township Key
1 Benoit
2 Melba
3 Bisley
4 Clifford
5 Ben Nevis
6 Pontiac
7 Maisonville
8 Bernhardt
9 Morrisette
10 Arnold
11 Katrine
12 Ossian
13 Grenfell
14 Teck
15 Lebel
16 Gauthier
17 McVittie
18 McGarry
19 Eby
20 Otto
21 Boston
22 McElroy
23 Hearst
24 Marquis
25 Pacaud
26 Catharine
27 Skead
28 Marter
29 Bayly
Kirkland Lake
Larder Lake
N
0
10 km
48°15'N
48°00'N
80°00'W
79°30'W
Proterozoic rocks
Felsic intrusive rocks
Alkalic metavolcanic rocks and associated metasedimentary rocks; minor ultramafic intrusive rocks
Calc-alkalic metavolcanic rocks
Tholeiitic and komatiitic metavolcanic rocks
Diabase and granophyre sheets and dikes
Geological boundary
Boundary of study area
Township boundary
Sample containing spinel
b

**Figure 27.11b. Geological maps of the Kirkland Lake Initiatives Program study area displaying locations of till samples containing spinel (modified *from* Fortescue and Gleeson 1984).**

## AREA GEOCHEMICAL PROVINCE PATTERNS

Several examples of regional area geochemical province patterns have already been described (e.g., Figures 27.9a, 27.9b, 27.13, 27.14, 27.15 and 27.16). A good example of a combined geochemical province–geochemical anomaly pattern occurs in the molybdenum data from the Hanes Lake map area (Figure 27.20; Fortescue and Vida 1990b). The mean Mo level in the Hanes Lake area is 1.5 Kk (1.85 ppm). The Mo geochemical province is delineated by values greater than or equal to 5 Kk (6.0 ppm) in an area of complex geology (Figure 27.20b). Two Mo anomalies with values of 34 Kk (41 ppm) and 29 Kk (35 ppm), respectively, occur within this geochemical province. At the time the Hanes Lake geochemical map was released, neither the molybdenum province nor the anomalies had been staked.

Area geochemical province patterns are not always due to bedrock geology. For example, a geochemical province for pH and for calcium in lakewater occurs in the west-central part of the Hanes Lake map (Figure 27.21; Fortescue and Vida 1990b). This geochemical province is associated with an area of calcareous ground moraine and outwash plain (*see* Figure 27.21). It is interesting that a geochemical province for arsenic (Figure 27.22b) is coincident with the pH and calcium patterns just described. The G-GMS geochemical province patterns for arsenic (Figure 27.22c, parts

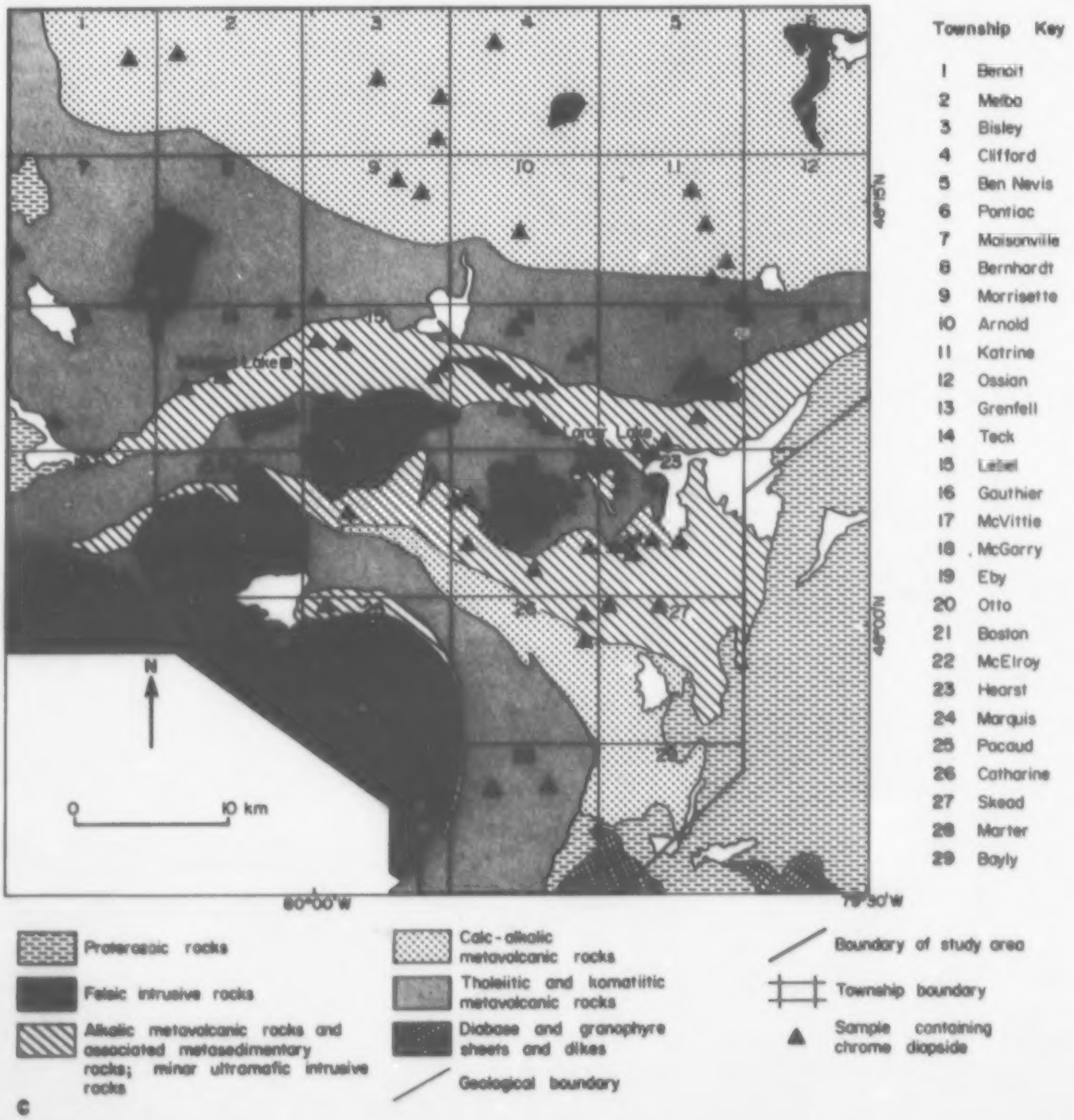

Township Key
1 Benoit
2 Melba
3 Bisley
4 Clifford
5 Ben Nevis
6 Pontiac
7 Maisonville
8 Bernhardt
9 Morrisette
10 Arnold
11 Katrine
12 Ossian
13 Grenfell
14 Teck
15 Lebel
16 Gauthier
17 McVittie
18 McGarry
19 Eby
20 Otto
21 Boston
22 McElroy
23 Hearst
24 Marquis
25 Pacaud
26 Catharine
27 Skead
28 Marter
29 Bayly
Kirkland Lake
Larder Lake
N
0
10 km
48°15'N
48°00'N
80°00'W
79°30'W
Proterozoic rocks
Felsic intrusive rocks
Alkalic metavolcanic rocks and associated metasedimentary rocks; minor ultramafic intrusive rocks
Calc-alkalic metavolcanic rocks
Tholeiitic and komatiitic metavolcanic rocks
Diabase and granophyre sheets and dikes
Geological boundary
Boundary of study area
Township boundary
Sample containing chrome diopside
c

**Figure 27.11c.** Geological maps of the Kirkland Lake Initiatives Program study area displaying locations of till samples containing chrome diopside (*modified from* Fortescue and Gleeson 1984).

C and D and parts B and C) show a strong geochemical gradient from north to south. Variation in the thickness of the ground moraine may account for this gradient. Also, the arsenic G-GMS includes a single anomalous value of 53 Kk (96 ppm) in Figure 27.22c, parts B and D, which has not yet been verified or explained.

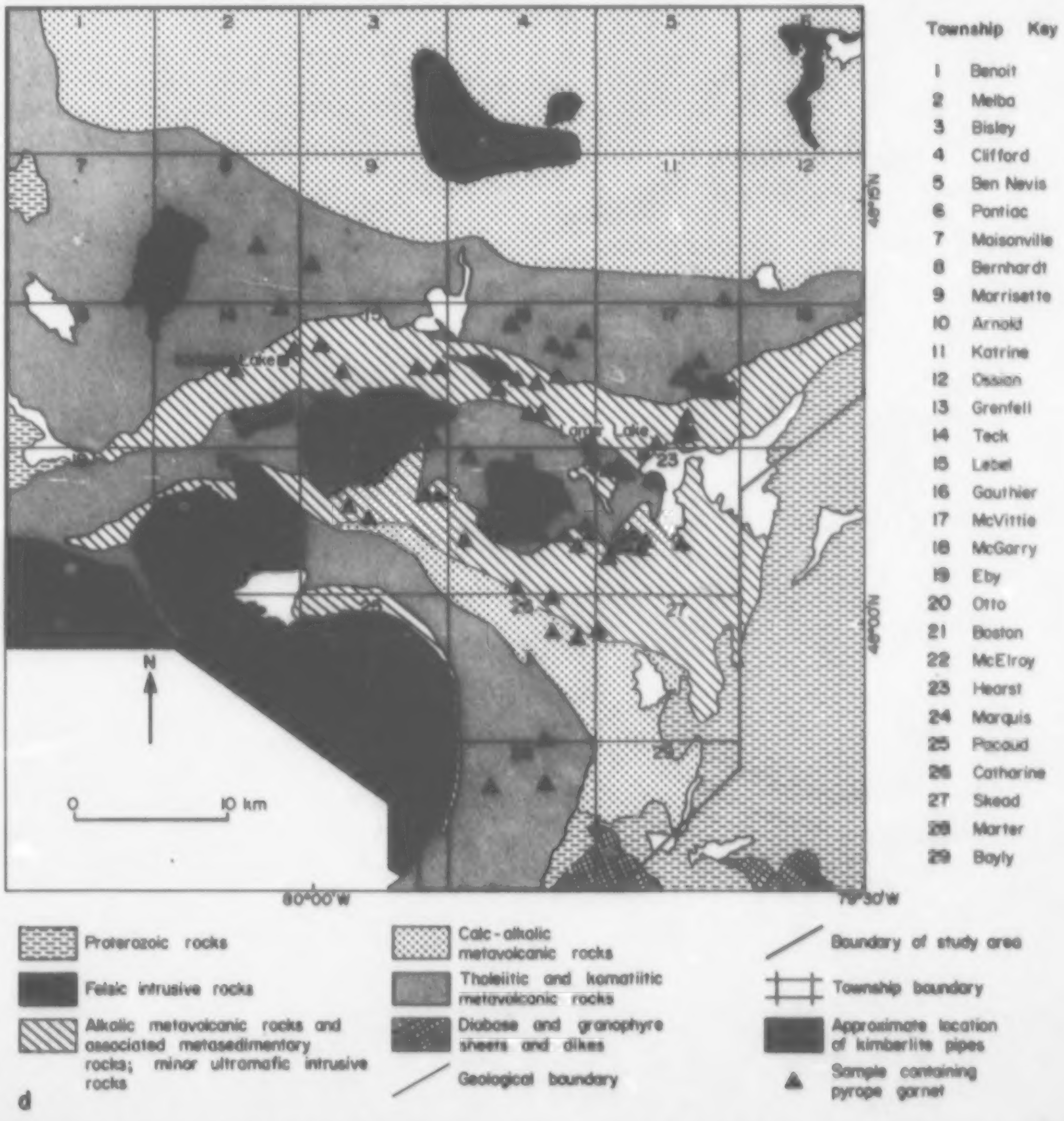
## COMMENTS

These examples of regional geochemical provinces, and many others like them, provide an introduction to pattern recognition in geochemical mapping. As more interpretive experience of this type is gained, it should be possible to predict the presence of some of these geochemical province patterns. As more experience is gained with GIS, it is probable that geochemical province patterns will be associated with similar patterns in OGS remote sensing and metallogeny data.

# PART IV: GEOCHEMICAL MAPPING RESEARCH AND DEVELOPMENT

## Geochemical Data Verification Methodology

Ideally, all geochemical anomalies and some geochemical province patterns should be verified by resampling soon after discovery. Unfortunately, this is not possible unless the program is serviced by a Mobile Laboratory Unit (MLU; *see* The Dedication of a Chemical Laboratory to Ontario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6
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12
14
15
17
18
Kirkland Lake
Larder Lake
23
26
27
28
29
48°15'N
48°00'N
80°00'W
79°30'W
N
0
10 km
Township Key
1 Benoit
2 Melba
3 Bisley
4 Clifford
5 Ben Nevis
6 Pontiac
7 Maisonville
8 Bernhardt
9 Morrisette
10 Arnold
11 Katrine
12 Ossian
13 Grenfell
14 Teck
15 Lebel
16 Gauthier
17 McVittie
18 McGarry
19 Eby
20 Otto
21 Boston
22 McElroy
23 Hearst
24 Marquis
25 Pacaud
26 Catharine
27 Skead
28 Marter
29 Bayly
Proterozoic rocks
Felsic intrusive rocks
Alkalic metavolcanic rocks and associated metasedimentary rocks; minor ultramafic intrusive rocks
Calc-alkalic metavolcanic rocks
Tholeiitic and komatiitic metavolcanic rocks
Diabase and granophyre sheets and dikes
Geological boundary
Boundary of study area
Township boundary
Approximate location of kimberlite pipes
Sample containing pyrope garnet
d

**Figure 27.11d.** Geological maps of the Kirkland Lake Initiatives Program study area displaying locations of kimberlite pipes and locations of till samples containing pyrope garnet (*modified from* Fortescue and Gleeson 1984).

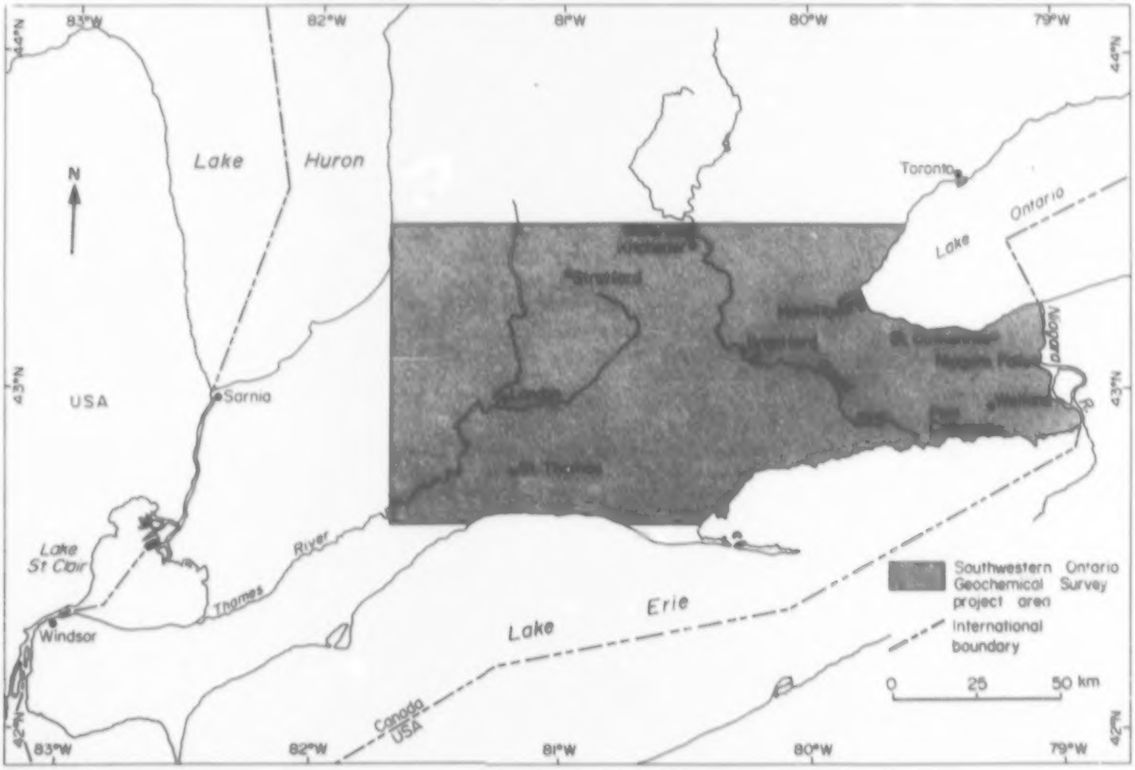
Geological Survey Mapping). Instead, the OGS has focussed attention on the development of resampling methods for verification of pre–*Ambrosia* lake sediment core anomaly data. The 4 approaches to the solution of this problem that have been examined are: 1) duplicate chemical analyses of the original pre–*Ambrosia* material; 2) chemical analysis of both pre– and post–*Ambrosia* samples from anomalous cores; 3) replicated sampling of pre–*Ambrosia* material from lakes of interest; and 4) the collection of a long lake sediment core from a sample point of interest (Fortescue 1988b).

Because of the excellent reliability of modern chemical analyses, approach 1 listed above is not usually necessary. For example, in recent OGS experience involving the preparation of 5 regional geochemical maps, only 2 minor analytical errors were found, both of which were easy to identify since they involved only a small number of elements.

An example of approach 2 along a 75 km geochemical sampling strip is illustrated on Figure 27.5. In this case, the pre– and post–*Ambrosia* patterns for strontium and uranium were very similar and unaffected by atmospheric fallout (*see* Figures 27.5d and 27.5f). The patterns for lead and nickel in the same samples (*see* Figures 27.5a and 27.5b) are dissimilar due to anthropogenic fallout from the atmosphere during the post–*Ambrosia* period. Consequently, this approach to anomaly verification in lake sediment cores cannot be used for all elements.

An example of approach 3 is the resampling of the pre–*Ambrosia* material to verify the zinc and copper anomaly shown in Figures 27.8b and 27.8c. Geochemical results obtained from 1988 and 1989 samplings from the 3 lakes of interest (*see* Figure 27.8) are included in Table 27.4. The data in Table 27.4 include: 1) levels for a global datum signature (GDS) based on the median Kk values for 32 elements, in 3000 pre–*Ambrosia* lake sediment samples (Fortescue and Vida, in prep.); 2) a list of Kk values for 32 elements in the 1988 sampling of lake sites EG599, EG603 and EG617; and 3) a list of Kk values for a 1989 sampling of lake sites EG599 (X1054), EG603 (X1055) and EG617 (X1056). Values for 5 important elements (Zn, Cu, As, Fe and Mn) are underlined on Table 27.4.

The contrast for the Zn geochemical anomaly is between 5.7 and 12.9, with a GDS value of 1.00 Kk (*see* Table 27.4). The contrasts for the Cu geochemical anomaly ranged between 1.8 and 4.1 with a GDS level of 0.49 Kk (*see* Table 27.4). Arsenic is another element associated with this geochemical anomaly (*see* Table 27.4). In the 1989 sampling, in 2 of the 3 samples of interest, the As values were close to the GDS value of 1.11 Kk. In the 1988 sampling, all 3 samples were found to be anomalous for As, with a contrast between 2.8 and 7.8 (*see* Table 27.4).


83°W
82°W
81°W
80°W
79°W
44°N
43°N
42°N
N
Lake
Huron
Toronto
Lake
Ontario
Kitchener
Stratford
Hamilton
Brantford
St. Catharines
Niagara Falls
Niagara R.
Welland
London
St. Thomas
USA
Sarnia
Lake St Clair
Windsor
Thames
River
Lake
Erie
Canada
USA
Southwestern Ontario Geochemical Survey project area
International boundary
0
25
50 km

a

**Figure 27.12a.** The Southwestern Ontario Geochemical Survey project area: location map (*modified from* Fortescue 1983a, 1984a).

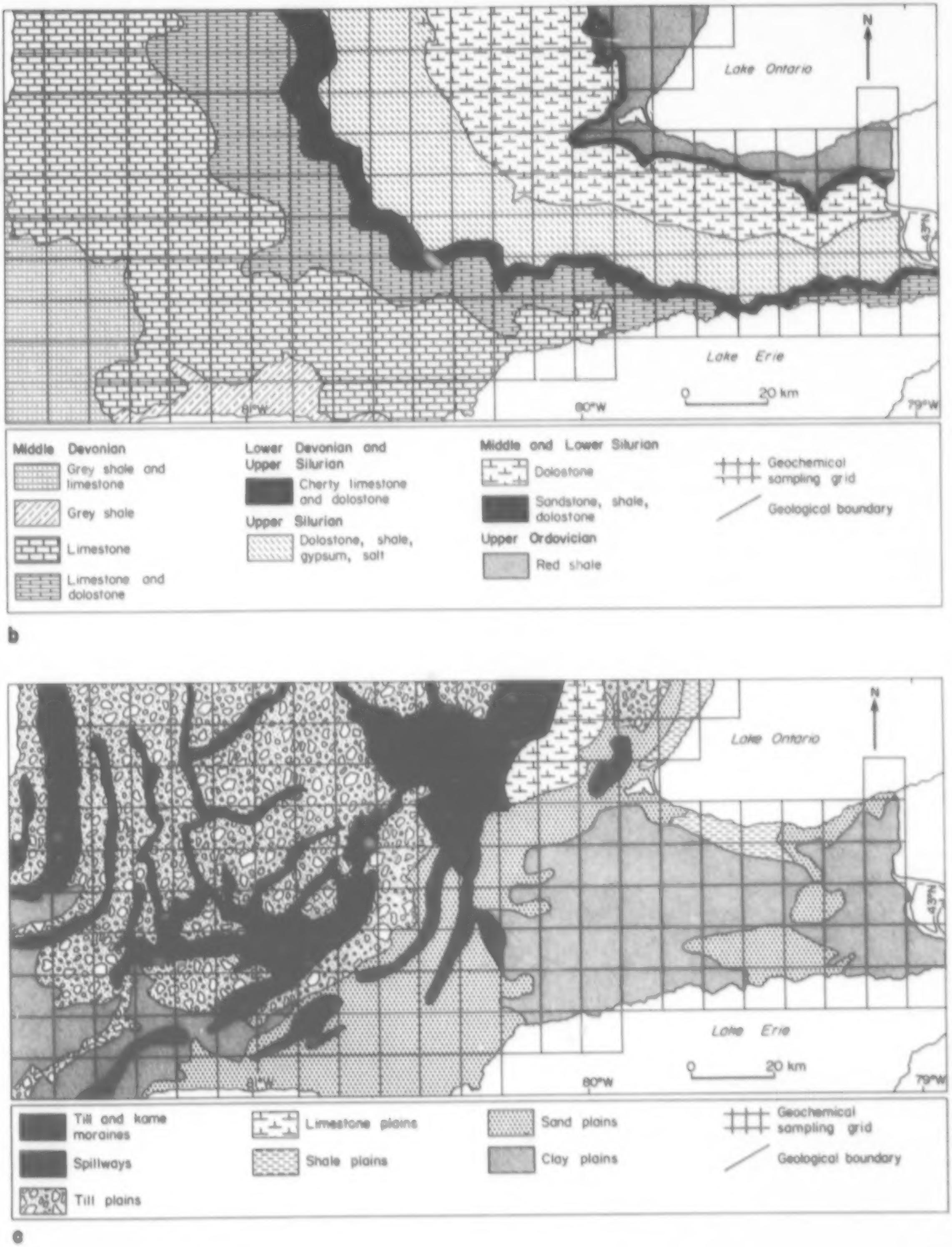

Lake Ontario
Lake Erie
N
0 20 km
81°W
80°W
79°W
43°N
Middle Devonian
Grey shale and limestone
Grey shale
Limestone
Limestone and dolostone
Lower Devonian and Upper Silurian
Cherty limestone and dolostone
Upper Silurian
Dolostone, shale, gypsum, salt
Middle and Lower Silurian
Dolostone
Sandstone, shale, dolostone
Upper Ordovician
Red shale
Geochemical sampling grid
Geological boundary
b
Till and kame moraines
Spillways
Till plains
Limestone plains
Shale plains
Sand plains
Clay plains
Geochemical sampling grid
Geological boundary
c

**Figure 27.12b and c.** The Southwestern Ontario Geochemical Survey project area: b) bedrock geology map; and c) Quaternary geology map (*modified from* Fortescue 1983a, 1984a).

At 2 of the 3 anomalous sample sites (EG603 and EG617), iron and manganese values were replicated well, at approximately half of their GDS value of 0.12 Kk (*see* Table 27.4). In the 2 samples from the headwater lake (EG599), values averaged 1.4 times the GDS value for iron and 1.9 times the GDS value for manganese. This suggests that the source of the geochemical anomaly may be gossan in the lake or catchment area of sample EG599. The data on Table 27.4 illustrates the importance of verification by resampling of pre-*Ambrosia* lake sediment and also illustrates how additional geochemical information can be obtained from this procedure.

Approach 4 is the most effective and sure method for verification of pre-*Ambrosia* geochemical data. This method involves the collection of long lake sediment cores. In central Ontario, the organic layer sampled by long lake sediment cores has accumulated during a 9000-year period

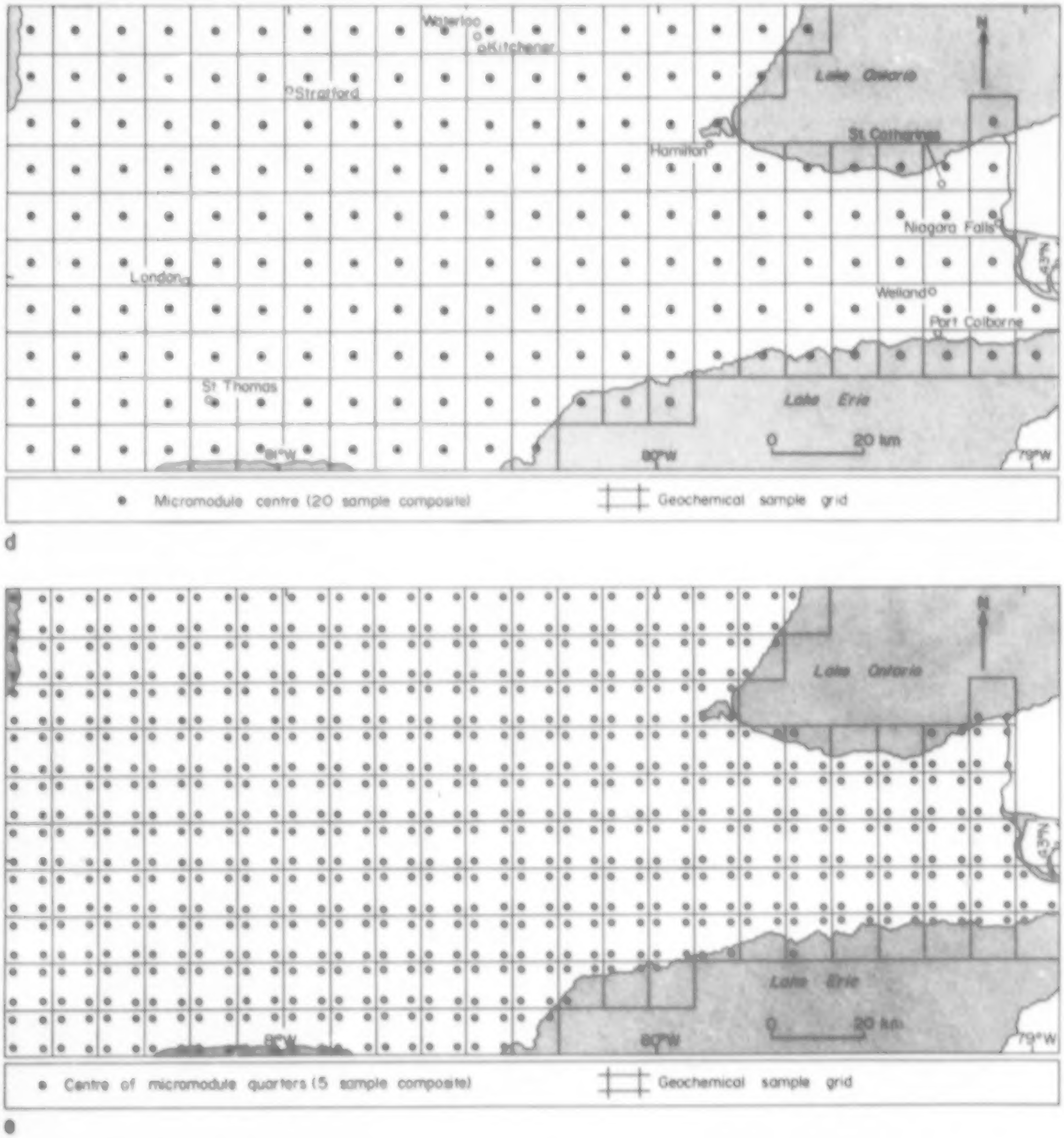
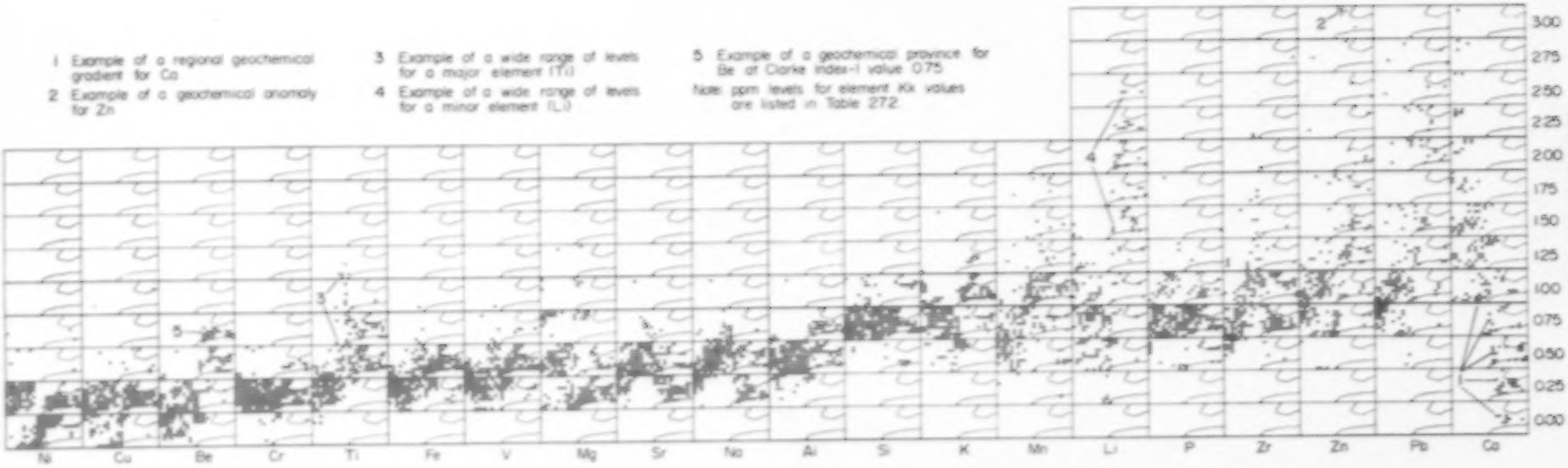

Waterloo
Kitchener
Stratford
Lake Ontario
N
St Catharines
Hamilton
Niagara Falls
London
43°N
Welland
Port Colborne
St Thomas
Lake Erie
0
20 km
81°W
80°W
79°W
Micromodule centre (20 sample composite)
Geochemical sample grid
d
Centre of micromodule quarters (5 sample composite)
Geochemical sample grid
e

**Figure 27.12d and e.** The Southwestern Ontario Geochemical Survey project area: **d)** map overprinted with the micromodule (20 sample) composite boundaries; and **e)** map overprinted with micromodule quarter (5 sample) composite boundaries. One micromodule consists of an area of 10 by 10 km and one micromodule quarter consists of an area of 5 by 5 km. For a more detailed explanation *see* text (*modified from* Fortescue 1983a, 1984a).



1 Example of a regional geochemical gradient for Ca

2 Example of a geochemical anomaly for Zn

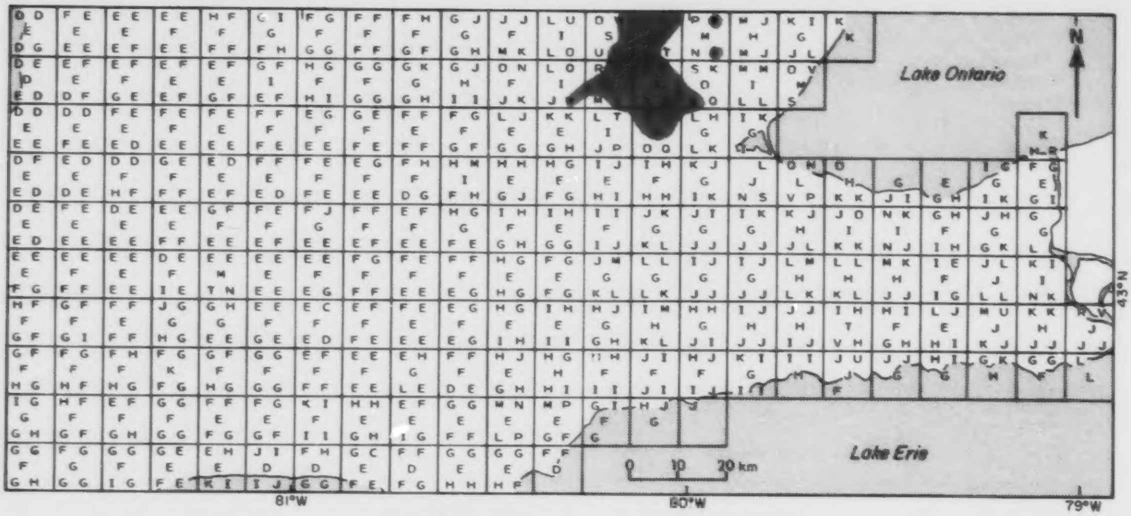
3 Example of a wide range of levels for a major element (Ti)

4 Example of a wide range of levels for a minor element (Li)

5 Example of a geochemical province for Be at Clarke index-1 value 0.75

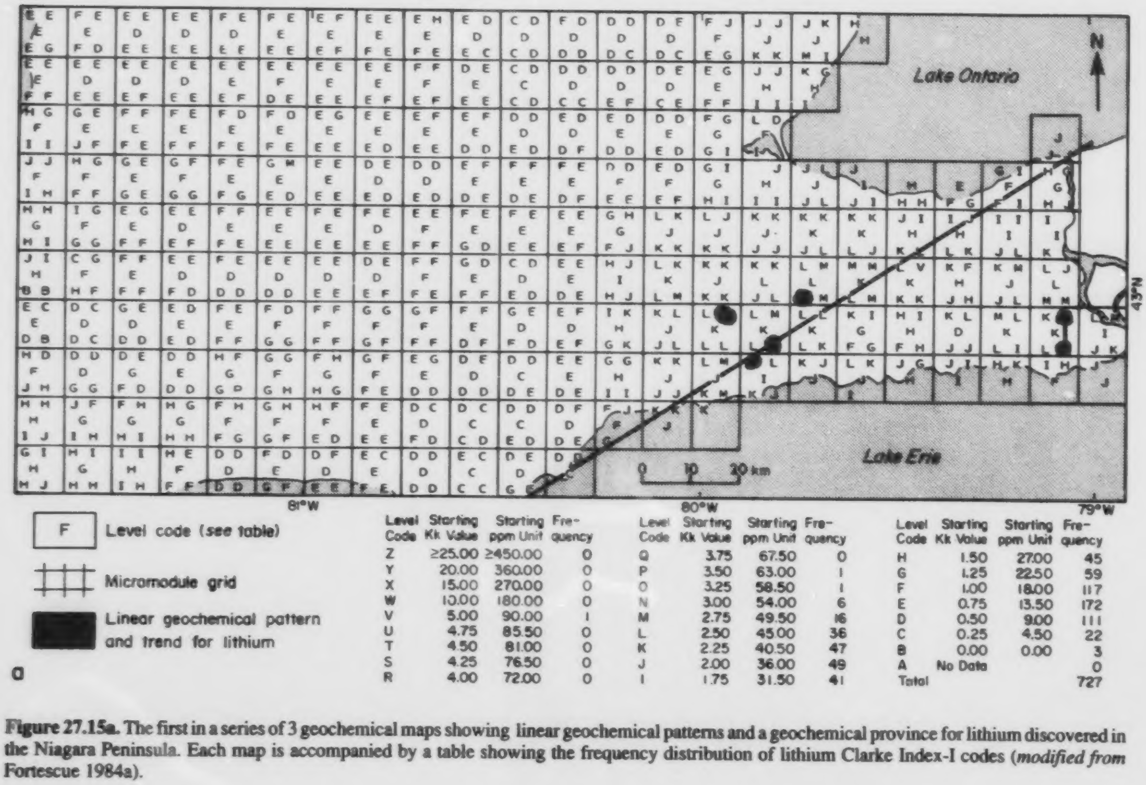
Note: ppm levels for element Kx values are listed in Table 27.2

**Figure 27.13.** Composite scan sheet summarizing regional geochemical patterns for 20 elements in over 4000 samples included in the Southwestern Ontario Geochemical Survey project (*modified from* Fortescue 1983a).


N
Lake Ontario
Lake Erie
0 10 20 km
81°W
80°W
79°W
43°N
F Level code (see table)
Micromodule grid
Geochemical anomaly for zinc

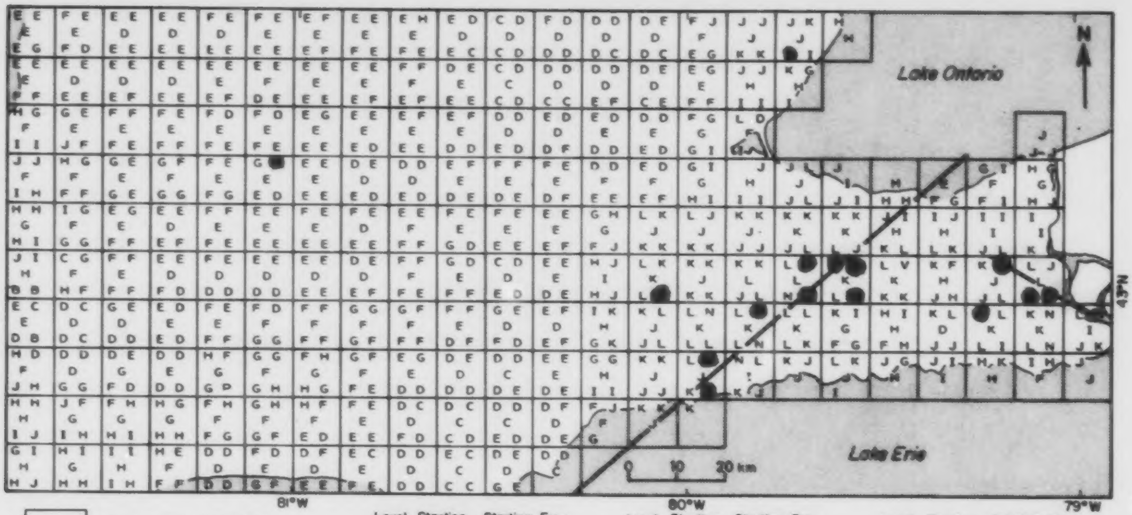
| Level Code | Starting Kk Value | Starting ppm Unit | Frequency |
|---|---|---|---|
| Z | ≥25.00 | ≥1900.00 | 1 |
| Y | 20.00 | 1520.00 | 1 |
| X | 15.00 | 1140.00 | 6 |
| W | 10.00 | 760.00 | 3 |
| V | 5.00 | 380.00 | 10 |
| U | 4.75 | 361.00 | 4 |
| T | 4.50 | 342.00 | 3 |
| S | 4.25 | 323.00 | 3 |
| R | 4.00 | 304.00 | 3 |
| Q | 3.75 | 285.00 | 4 |
| P | 3.50 | 266.00 | 5 |
| O | 3.25 | 247.00 | 7 |
| N | 3.00 | 228.00 | 9 |
| M | 2.75 | 209.00 | 14 |
| L | 2.50 | 190.00 | 31 |
| K | 2.25 | 171.00 | 38 |
| J | 2.00 | 152.00 | 65 |
| I | 1.75 | 133.00 | 63 |
| H | 1.50 | 114.00 | 73 |
| G | 1.25 | 95.00 | 118 |
| F | 1.00 | 76.00 | 117 |
| E | 0.75 | 57.00 | 12 |
| D | 0.50 | 38.00 | 26 |
| C | 0.25 | 19.00 | 2 |
| B | 0.00 | 0.00 | 0 |
| A | No Data | | 0 |
| Total | | | 727 |

**Figure 27.14.** Map showing a regional geochemical anomaly for zinc in the Southwestern Ontario Geochemical Survey area. The accompanying table shows frequency distribution of zinc Clarke Index-I codes used in the preparation of the map (*modified from* Fortescue 1984a).


N
Lake Ontario
Lake Erie
0 10 20 km
81°W
80°W
79°W
43°N
F Level code (see table)
Micromodule grid
Linear geochemical pattern and trend for lithium
a

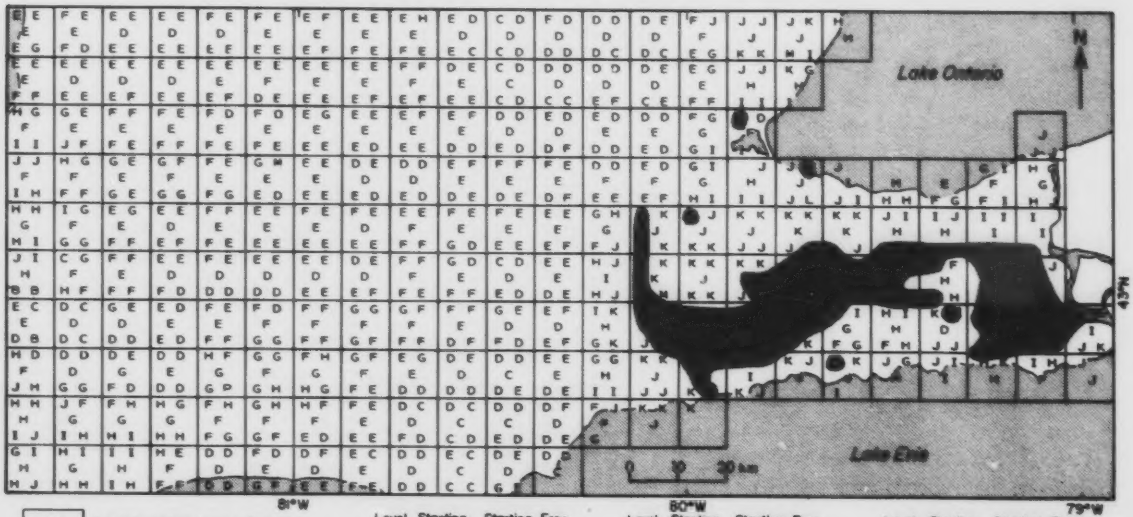
| Level Code | Starting Kk Value | Starting ppm Unit | Frequency |
|---|---|---|---|
| Z | ≥25.00 | ≥450.00 | 0 |
| Y | 20.00 | 360.00 | 0 |
| X | 15.00 | 270.00 | 0 |
| W | 10.00 | 180.00 | 0 |
| V | 5.00 | 90.00 | 1 |
| U | 4.75 | 85.50 | 0 |
| T | 4.50 | 81.00 | 0 |
| S | 4.25 | 76.50 | 0 |
| R | 4.00 | 72.00 | 0 |
| Q | 3.75 | 67.50 | 0 |
| P | 3.50 | 63.00 | 1 |
| O | 3.25 | 58.50 | 1 |
| N | 3.00 | 54.00 | 6 |
| M | 2.75 | 49.50 | 16 |
| L | 2.50 | 45.00 | 36 |
| K | 2.25 | 40.50 | 47 |
| J | 2.00 | 36.00 | 49 |
| I | 1.75 | 31.50 | 41 |
| H | 1.50 | 27.00 | 45 |
| G | 1.25 | 22.50 | 59 |
| F | 1.00 | 18.00 | 117 |
| E | 0.75 | 13.50 | 172 |
| D | 0.50 | 9.00 | 111 |
| C | 0.25 | 4.50 | 22 |
| B | 0.00 | 0.00 | 3 |
| A | No Data | | 0 |
| Total | | | 727 |

**Figure 27.15a.** The first in a series of 3 geochemical maps showing linear geochemical patterns and a geochemical province for lithium discovered in the Niagara Peninsula. Each map is accompanied by a table showing the frequency distribution of lithium Clarke Index-I codes (*modified from* Fortescue 1984a).


Lake Ontario
Lake Erie
N
0 10 20 km
81°W
80°W
79°W
43°N
F Level code (*see* table)
Micromodule grid
Linear geochemical pattern and trend for lithium
b

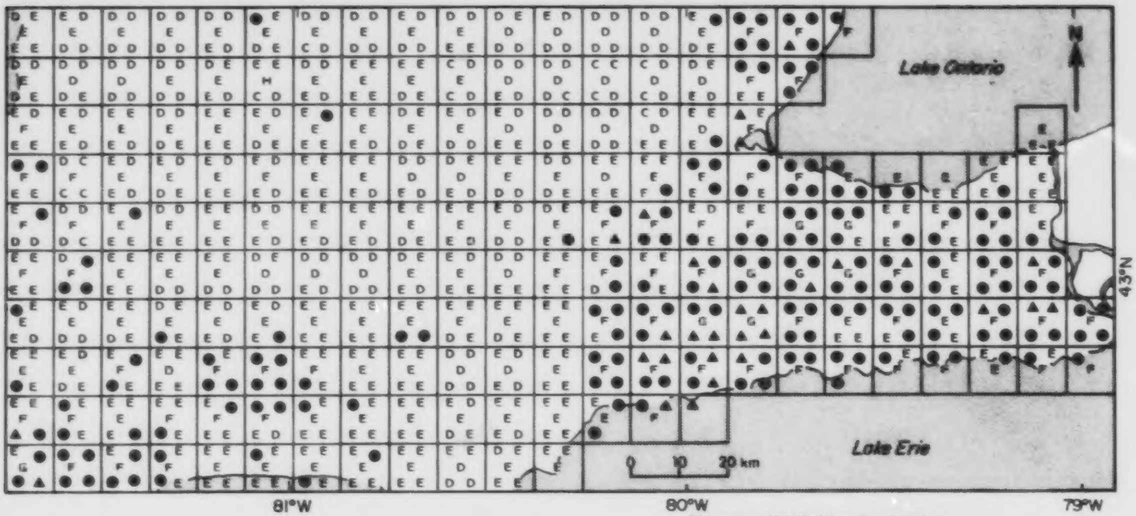
| Level Code | Starting Kk Value | Starting ppm Unit | Frequency | Level Code | Starting Kk Value | Starting ppm Unit | Frequency | Level Code | Starting Kk Value | Starting ppm Unit | Frequency |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Z | ≥25.00 | ≥450.00 | 0 | Q | 3.75 | 67.50 | 0 | H | 1.50 | 27.00 | 45 |
| Y | 20.00 | 360.00 | 0 | P | 3.50 | 63.00 | 1 | G | 1.25 | 22.50 | 59 |
| X | 15.00 | 270.00 | 0 | O | 3.25 | 58.50 | 1 | F | 1.00 | 18.00 | 117 |
| W | 10.00 | 180.00 | 0 | N | 3.00 | 54.00 | 6 | E | 0.75 | 13.50 | 172 |
| V | 5.00 | 90.00 | 1 | M | 2.75 | 49.50 | 16 | D | 0.50 | 9.00 | 111 |
| U | 4.75 | 85.50 | 0 | L | 2.50 | 45.00 | 36 | C | 0.25 | 4.50 | 22 |
| T | 4.50 | 81.00 | 0 | K | 2.25 | 40.50 | 47 | B | 0.00 | 0.00 | 3 |
| S | 4.25 | 76.50 | 0 | J | 2.00 | 36.00 | 49 | A | No Data | | 0 |
| R | 4.00 | 72.00 | 0 | I | 1.75 | 31.50 | 41 | Total | | | 727 |

**Figure 27.15b.** The second in a series of 3 geochemical maps showing linear geochemical patterns and a geochemical province for lithium discovered in the Niagara Peninsula. Each map is accompanied by a table showing the frequency distribution of lithium Clarke Index-I codes (*modified from* Fortescue 1984a).


Lake Ontario
Lake Erie
N
0 10 20 km
81°W
80°W
79°W
43°N
F Level codes (*see* table)
Micromodule grid
Geochemical province for lithium
c

| Level Code | Starting Kk Value | Starting ppm Unit | Frequency | Level Code | Starting Kk Value | Starting ppm Unit | Frequency | Level Code | Starting Kk Value | Starting ppm Unit | Frequency |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Z | ≥25.00 | ≥450.00 | 0 | Q | 3.75 | 67.50 | 0 | H | 1.50 | 27.00 | 45 |
| Y | 20.00 | 360.00 | 0 | P | 3.50 | 63.00 | 1 | G | 1.25 | 22.50 | 59 |
| X | 15.00 | 270.00 | 0 | O | 3.25 | 58.50 | 1 | F | 1.00 | 18.00 | 117 |
| W | 10.00 | 180.00 | 0 | N | 3.00 | 54.00 | 6 | E | 0.75 | 13.50 | 172 |
| V | 5.00 | 90.00 | 1 | M | 2.75 | 49.50 | 16 | D | 0.50 | 9.00 | 111 |
| U | 4.75 | 85.50 | 0 | L | 2.50 | 45.00 | 36 | C | 0.25 | 4.50 | 22 |
| T | 4.50 | 81.00 | 0 | K | 2.25 | 40.50 | 47 | B | 0.00 | 0.00 | 3 |
| S | 4.25 | 76.50 | 0 | J | 2.00 | 36.00 | 49 | A | No Data | | 0 |
| R | 4.00 | 72.00 | 0 | I | 1.75 | 31.50 | 41 | Total | | | 727 |

**Figure 27.15c.** The third in a series of 3 geochemical maps showing linear geochemical patterns and a geochemical province for lithium discovered in the Niagara Peninsula. The map is accompanied by a table showing the frequency distribution of lithium Clarke Index-I codes (*modified from* Fortescue 1984a).

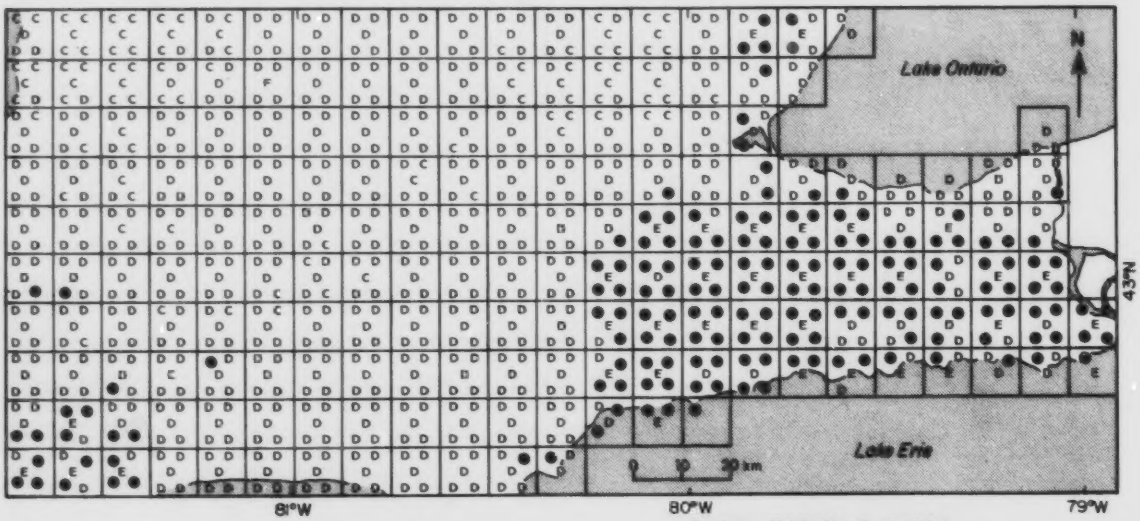


| | |
|---|---|
| F | Level code (*see table*) |
| ┼┼┼ | Micromodule grid |
| ● | Geochemical anomaly for level code F |
| ▲ | Geochemical anomaly for level code G |

a

Frequency distribution of potassium

| Level Code | Starting Kk Value | Starting ppm Unit | Fre-quency | Phase II Symbols |
|---|---|---|---|---|
| H | 1.50 | 27 600.00 | 0 | |
| G | 1.25 | 23 000.00 | 30 | ▲ |
| F | 1.00 | 18 400.00 | 184 | ● |
| E | 0.75 | 13 800.00 | 312 | |
| D | 0.50 | 9 200.00 | 186 | |
| C | 0.25 | 4 600.00 | 15 | |
| B | 0.00 | 0.00 | 0 | |
| A | No Data | | 0 | |
| Total | | | 727 | |



| | |
|---|---|
| F | Level code (*see table*) |
| ┼┼┼ | Micromodule grid |
| ● | Geochemical anomaly (level codes F,E) |

b

Frequency distribution of aluminium

| Level Code | Starting Kk Value | Starting ppm Unit | Fre-quency | Phase II Symbols |
|---|---|---|---|---|
| H | 1.50 | 125 400.00 | 0 | |
| G | 1.25 | 104 500.00 | 0 | |
| F | 1.00 | 83 600.00 | 2 | ● |
| E | 0.75 | 62 700.00 | 166 | ● |
| D | 0.50 | 41 800.00 | 489 | |
| C | 0.25 | 20 900.00 | 70 | |
| B | 0.00 | 0.00 | 0 | |
| A | No Data | | 0 | |
| Total | | | 727 | |

**Figure 27.16a and b.** Geochemical maps of the entire Southwestern Ontario Geochemical Survey area showing geochemical province patterns for: **a)** potassium (K); and **b)** aluminum (Al). Each map is accompanied by a table showing the frequency distribution of Clarke Index-I values for each element (*modified from* Fortescue 1984a).

(Saarnisto 1975). These cores, which are usually from 3 to 5 m long, extend from the bottom of a lake through the entire layer of organic sediment into the underlying mineral matter.

For example, Fortescue and Vida (1990a) described a cluster of 5 anomalous copper values in pre–*Ambrosia* short core samples from the Herman Lake regional geochemical survey (Figure 27.23). These cores were collected during the summer of 1987. In March 1988, using the vibrocorer drilling technique described by Fortescue (1988b), a long lake sediment core was taken at one of these points (core 24J is indicated by an arrow on Figure 27.23).

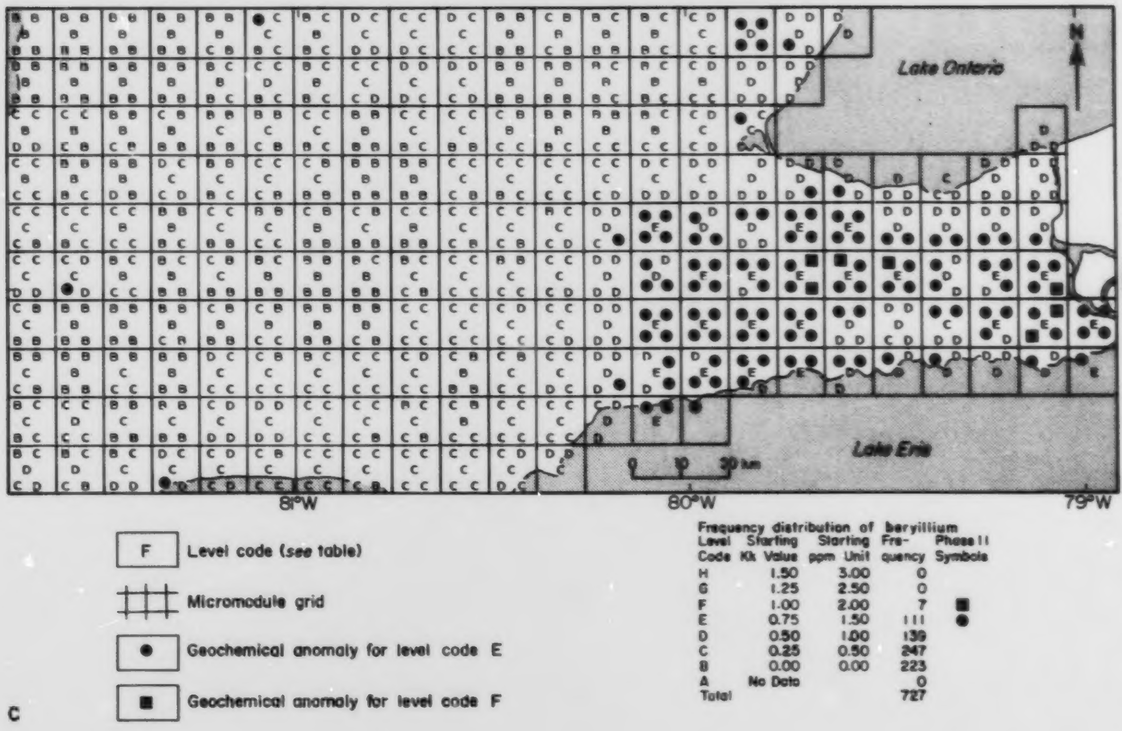
Table 27.5 lists, in full, the data for lanthanum, copper, molybdenum, gold and arsenic in long core 24J. The 1987 copper geochemical anomaly is verified by this data. Also, the 5 high copper values actually extend to the bottom [illegible] organic layer. The copper anomaly does not contin[illegible] the underlying mineral matter. This suggests that the source of the geochemical anomaly is not in the immediate vicinity of the drill hole. Table 27.5 also shows enhanced values for molybdenum, gold and arsenic in the lower organic part of the core. Like all long cores examined so far, core 24J has a marked enrichment of copper near the base of the organic layer. This is probably associated with an increase in *Betula* pollen (Fortescue 1986). At the time of writing, the lanthanum pattern in core 24J had not been explained.

## SUMMARY

Long cores are certainly the most effective of the 4 approaches to pre–*Ambrosia* lake sediment data verification. Long cores verify the location of geochemical anomalies and provide information on the duration of element input into sediments at a drill site.

## Environmental Geochemistry Research

Between 1980 and 1987, the OGS supported a small research program in environmental geochemistry (Fortescue et al. 1981; Fortescue, Dickman et al. 1984; Fortescue 1984b, 1986; Fortescue and Stahl 1987a, 1987b). Initially, the aim of the program was to investigate the relationships between geochemical mapping and effects of the acid rain in the Wawa area. Experience gained from this environmental geochemistry research contributed substantially to the development of the lake sediment regional geochemical mapping methodology (*see* Figure 27.6). For this reason, some additional results of the environmental research are summarized here.


Lake Ontario
N
Lake Erie
0 10 30 km
81°W
80°W
79°W
43°N
F Level code (*see* table)
Micromodule grid
Geochemical anomaly for level code E
Geochemical anomaly for level code F
c

Frequency distribution of beryillium

| Level Code | Starting Kk Value | Starting ppm Unit | Frequency | Phase II Symbols |
|---|---|---|---|---|
| H | 1.50 | 3.00 | 0 | |
| G | 1.25 | 2.50 | 0 | |
| F | 1.00 | 2.00 | 7 | ■ |
| E | 0.75 | 1.50 | 111 | ● |
| D | 0.50 | 1.00 | 139 | |
| C | 0.25 | 0.50 | 247 | |
| B | 0.00 | 0.00 | 223 | |
| A | No Data | | 0 | |
| Total | | | 727 | |

**Figure 27.16c.** Geochemical maps of the entire Southwestern Ontario Geochemical Survey area showing geochemical province patterns for beryllium (Be). The map is accompanied by a table showing the frequency distribution of Clarke Index-I values for each element (*modified from* Fortescue 1984a).

## WATER GEOCHEMISTRY ALONG A pH GRADIENT NEAR WAWA

The first experiment in the environmental program was a comparative study of the water geochemistry in a series of 20 lakes. The lakes were located at ~5 km intervals along a sampling strip, 100 (northeast-southwest) by 10 km (northwest-southeast), positioned west of Wawa and the Herman Lake map area (*see* Figure 27.17). The sampling strip was selected on the basis of pH data obtained from a prior reconnaissance level geochemical survey completed in 1978 (OGS–GSC 1979). The sampling strip was chosen to include 20 lakes within a pH range from 4.5 to 8.0 (Figure 27.24; Fortescue et al. 1981; Fortescue 1984b).

It was concluded from this study that: 1) the pattern of pH variation in the 20 lakes was almost identical from year to year (Fortescue 1984b); and 2) the pH of the 20 lakes was governed by a combination of the geochemistry of the bedrock and Quaternary deposits in their catchment areas.

Both results are important from the viewpoint of regional geochemical mapping. The former, because it indicated pH patterns on adjacent geochemical map sheets (compiled during a multiyear geochemical mapping program) would be realistic; and the latter, because it indicated that in the Wawa area, the geological setting of lakes would likely overprint lakewater pH patterns that were due to acid rain.

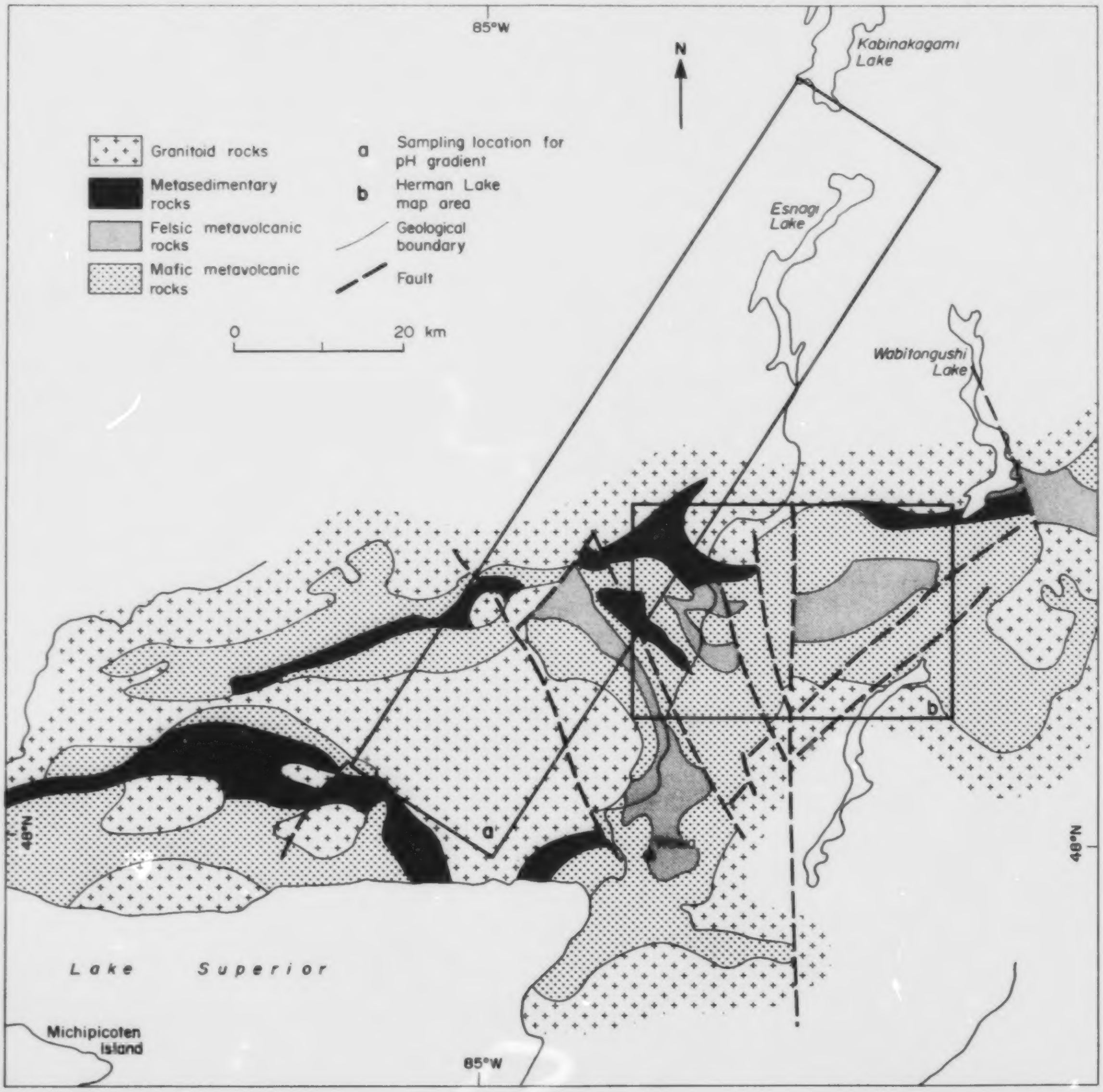

85°W
N
Kabinakagami Lake
Granitoid rocks
Metasedimentary rocks
Felsic metavolcanic rocks
Mafic metavolcanic rocks
a Sampling location for pH gradient
b Herman Lake map area
Geological boundary
Fault
0
20 km
Esnagi Lake
Wabitongushi Lake
a
b
48°N
48°N
Lake Superior
Michipicoten Island
85°W

**Figure 27.17.** Generalized geological map of the Michipicoten greenstone belt near Wawa showing: **a)** the location of the pH gradient selected for the acid rain research; and **b)** the extent of the Herman Lake regional geochemical map area (*modified from* Fortescue and Vida 1990a).

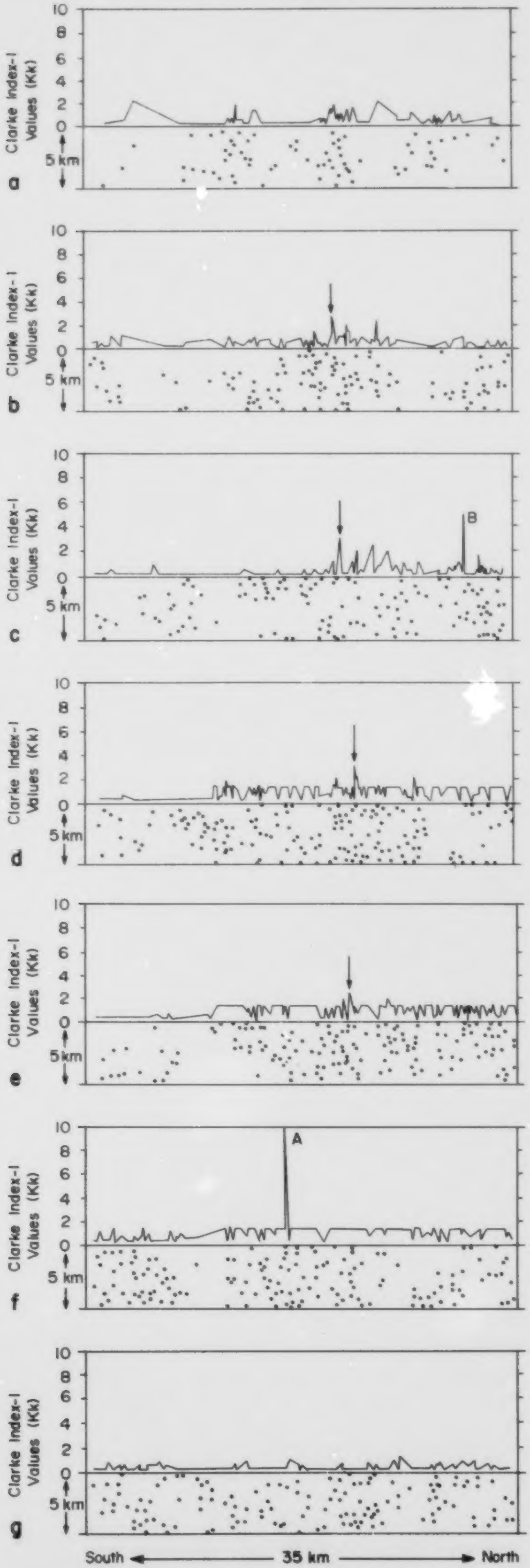


Figure 27.18. Geochemical section diagrams showing a linear geochemical province for gold in the Trout Lake regional geochemical map area. Unexplained isolated high gold values also occur at A and B. The black dots at the base of each section represent sample site locations. *See* Figure 27.8 for a map showing the location and general geology of the Trout Lake study area (*modified from* Fortescue and Vida 1989).

## DETERMINATION OF THE pH HISTORY OF WAWA AREA LAKES

The limnologist in the OGS interdisciplinary environmental geochemistry research team was Professor Dickman, of the Department of Biological Sciences at Brock University, St. Catharines, Ontario. During the acid rain research project, Dickman and his student researchers developed a standardized approach for the description of pH history of Wawa area lakes for the past 100 years (Dickman and Fortescue 1984, 1991; Fortescue, Dickman et al. 1984).

The standardized method was based on the detailed study of diatom remains in 1 cm sections of lake sediment cores 20 to 25 cm long. To estimate the pH of a particular core section, pH-sensitive diatoms were identified and population statistics obtained and combined; the result was then related to a working curve. The working curve was based on similar measurements of diatom floras taken from the bottom of lakes of known pH. Examples of graphs for the "diatom inferred pH" history of lakes during the past 100 years (based on this methodology) are provided for the Wawa area (Fortescue 1984b) and the Sudbury area (Dickman and Fortescue 1984).

This research showed that lakes and ponds in central Ontario may become acid due to: 1) a natural accumulation of humic acids; 2) forest fires in catchments (which change the pH of waters for a relatively short time); 3) chemical fallout from nearby smelters; and 4) non-point smelter fallout of acid rain derived from a distant source. It was noted that acid rain has the most drastic effect on nutrient-poor lakes which are usually located in catchments underlain by quartzites (Dickman and Fortescue 1984).

This research aided the development of regional geochemical mapping in several important ways: 1) it led to the development of a simple, rapid and reliable lake sediment core sampler (Fortescue 1988a); 2) it provided detailed information on the depth of the *Ambrosia* pollen rise in central Ontario lakes, leading to the standardization of pre–*Ambrosia* sediment as a sample medium for geochemical mapping; 3) it was shown that subsequent to the deposition of the sediment, no significant vertical movement of elements occurs in organic lake sediments, indicating that pre–*Ambrosia* material could be used with confidence to map pre–anthropogenic geochemical patterns; and 4) it was shown (Fortescue, Dickman et al. 1984; Fortescue 1988a) that in spite of bioturbation by small organisms, a lake sediment core can be dated using: a) pollen (e.g., the *Ambrosia* rise); and b) radio-isotopes ($^{210}$Pb, $^{137}$Cs etc.).

## SUMMARY

The acid rain environmental geochemistry program made substantial contributions towards the choice of pre-*Ambrosia* lake sediment as a medium for regional geochemical mapping by the OGS.

## The Use of Lake Sediments in Environmental Geochemistry

Simultaneous pre– and post-*Ambrosia* sampling during geochemical mapping is feasible using the lake sediment core sampling technique just discussed. If the 2 samples (i.e., pre– and post–*Ambrosia*) from each lake sediment core are analyzed separately, using identical methodologies (*see* Figure 27.6), the resulting element maps provide information on pre– and post–anthropogenic geochemical patterns similar to those described previously (*see* Figure 27.5).

The Herman Lake regional geochemical map area (Fortescue and Vida 1990a) is located in the Michipicoten greenstone belt some 40 km northeast of Wawa (*see* Figure 27.17). The Herman Lake regional geochemical map is a 1000 km² area, located 40 km downwind from the Wawa sintering plant—a point source of chemical fallout.

During the Herman Lake geochemical survey, 742 pre–*Ambrosia*, and 1 post–*Ambrosia* lake sediment cores were collected. Both sets of lake sediment samples were analyzed for 35 elements using the methodology summarized in Figure 27.6. The resulting geochemical data were used to prepare pre– and post–*Ambrosia* based regional geochemical maps for: 1) manganese and copper (control elements); 2) lead (non-point source pollution); 3) arsenic and antimony (point source pollution from the Wawa sintering plant); and 4) gold (element associated with prospecting and mining activity). These are shown in Figure 27.25 (Fortescue and Vida, in prep.).

Each of these geochemical maps includes: 1) an outline of the bedrock geology of the area; 2) lake sediment sample sites (small dots); and 3) lake sediment sample sites with values greater than or equal to an arbitrary threshold selected to display an anthropogenic pattern (large dots).

Briefly, the copper and manganese patterns were similar for both maps in each set. The pattern for the post–*Ambrosia* lead map (*see* Figure 27.25c) clearly shows an increased lead level due to atmospheric fallout from a non-point source. Post–*Ambrosia* patterns for arsenic and antimony are due to anthropogenic fallout of these elements from a point source (i.e., the Wawa sintering plant). Note that both these pre–*Ambrosia* element patterns include a geochemical province along the southern margin of the map. The post–*Ambrosia* pattern for gold shows 2 types of anthropogenic patterns (A and B) both associated with mining activity (*see* Figure 27.25f). The high post–*Ambrosia* gold values in the top right centre of the map (area A) are due to

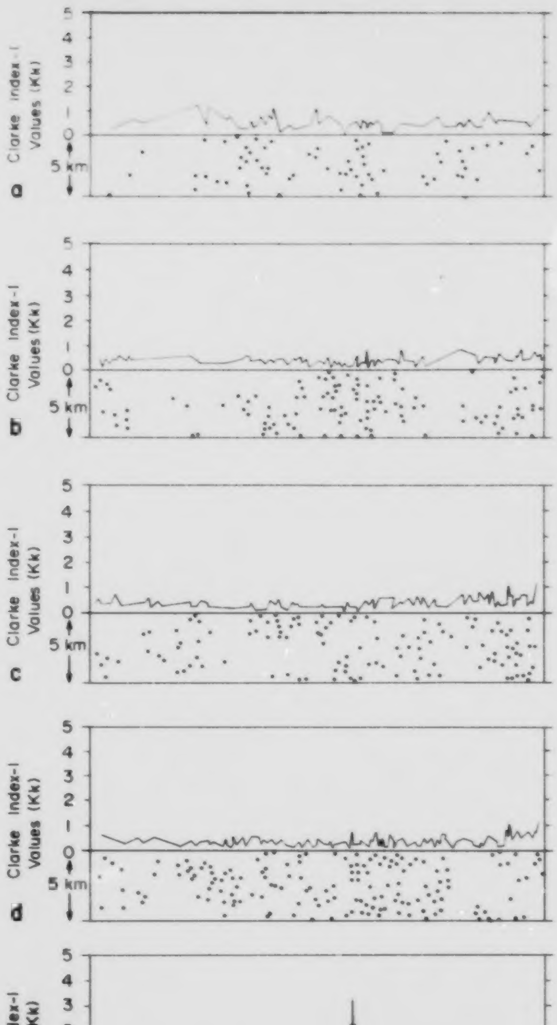


**Figure 27.19.** Geochemical section diagrams showing a linear geochemical province for thorium in the Trout Lake regional geochemical map area. The black dots at the base of each section represent sample site locations. *See* Figure 27.8 for a map showing the location and general geology of the Trout Lake study area (*modified from* Fortescue and Vida 1989).

surface disturbance of the landscape during recent prospecting activity. The area marked B is the site of an abandoned gold mine where tailings have contaminated lakes with small amounts of gold.

## SUMMARY

A regional geochemical mapping methodology based on separate pre– and post–*Ambrosia* based maps was demonstrated for the first time in the Wawa area. The map pairs are of interest in modern environmental geochemistry because they map: 1) effects of anthropogenic fallout; 2) contamination of the environment by mining activity; and 3) pre-anthropogenic patterns of interest in mineral resource appraisal and, possibly, forest nutrition.

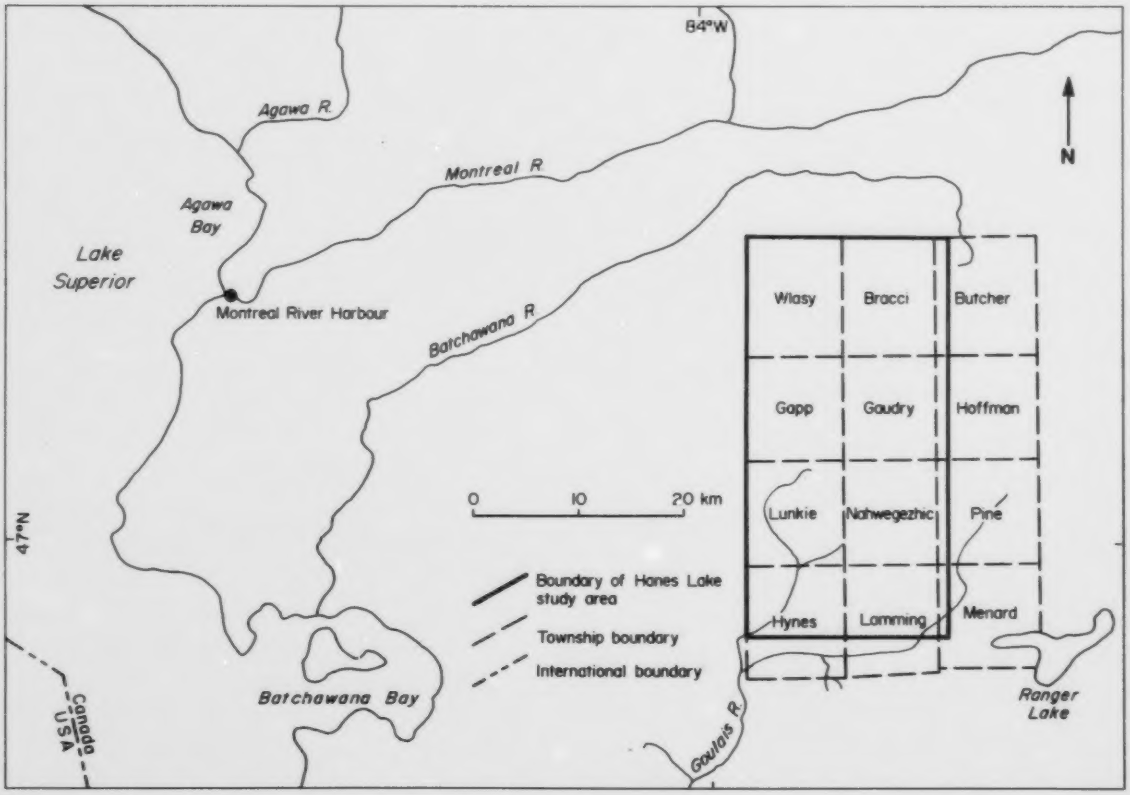
# PART V: DISCUSSION AND CONCLUSIONS

## General Discussion

Part I of this chapter introduced regional geochemical mapping and its role in the mineral resource appraisal process. This was followed by a description of how the geochemical data can be incorporated into a geoscience-oriented Geographic Information System. Part I also stressed the role played by regional geochemical mapping in the current Ontario Geological Survey approach to mineral resource appraisal of Ontario.

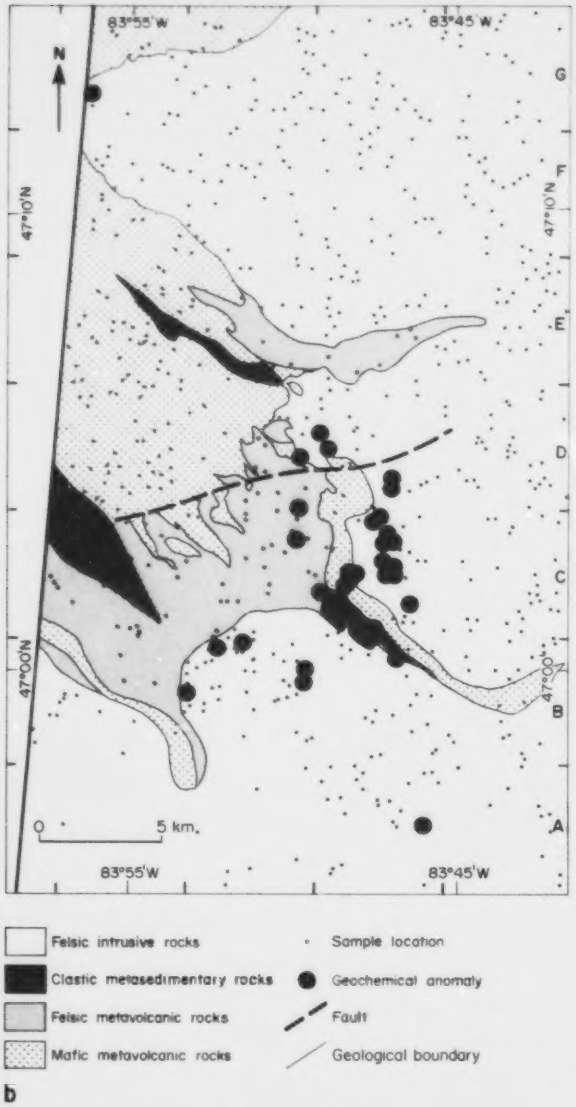
At the time of writing, the OGS regional geochemical mapping program is still being developed. For example, it is expected that the number of elements determined in each sample will rise from 35 to around 60 elements in the near future (Table 27.6). As experience is gained with the interpretation of geochemical map patterns for the new elements, the importance of the geochemical mapping component in an Ontario Geological Survey Geographic Information System should increase rapidly. When a similar situation occurred in the British Geological Survey some years ago, it led the British Geological Survey to officially recognize that geological, geophysical and geochemical maps are of equal importance in the mineral resource appraisal process (Jane Plant, British Geological Survey, personal communication, 1989).

Part II of this chapter described the process of regional geochemical mapping on the basis of 3 rules which currently guide the planning and interpretation of OGS regional geochemical maps. Examples taken from the OGS experience were used to illustrate the application of each rule. Major types of geochemical patterns currently


84°W
Agawa R.
N
Montreal R.
Agawa Bay
Lake Superior
Montreal River Harbour
Batchawana R.
Wlasy
Bracci
Butcher
Gapp
Gaudry
Hoffman
0
10
20 km
Lunkie
Nahwegezhic
Pine
47°N
Boundary of Hanes Lake study area
Hynes
Lamming
Menard
Township boundary
International boundary
Batchawana Bay
Ranger Lake
Canada
USA
Goulais R.

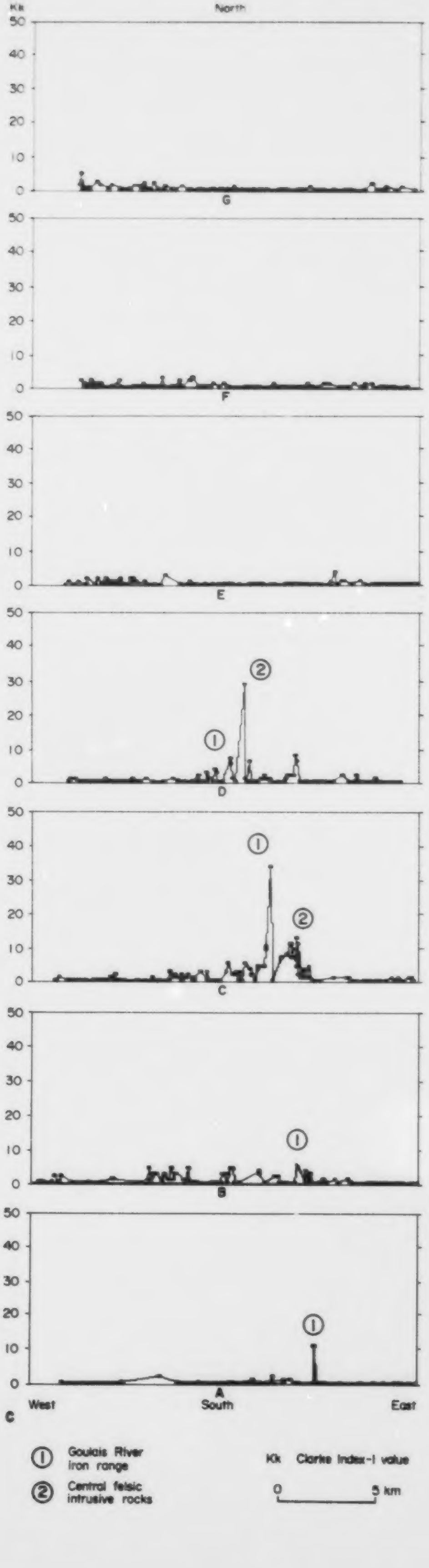
a

**Figure 27.20a.** Map showing the location of the Hanes Lake map area (*modified from* Fortescue and Vida 1990b).


83°55'W
83°45'W
N
G
F
E
D
C
B
A
47°10'N
47°00'N
0
5 km.
Felsic intrusive rocks
Clastic metasedimentary rocks
Felsic metavolcanic rocks
Mafic metavolcanic rocks
Sample location
Geochemical anomaly
Fault
Geological boundary

**b**

**Figure 27.20b.** Map showing the location of geochemical sample sites and anomalies with respect to general geology of the Hanes Lake area. Geochemical anomalies represent Clarke Index-I values greater than or equal to 5.0 Kk for molybdenum (*modified from* Fortescue and Vida 1990b).



**c**

**Figure 27.20c.** Geochemical section diagrams showing combined molybdenum geochemical anomaly and province patterns in the Hanes Lake area. The sections are lettered to correspond with the appropriate area of the map in part **b** (*modified from* Fortescue and Vida 1990b).

identified on OGS regional geochemical maps were also described in this part of the chapter.

In Part III, results from the 3 approaches to regional geochemical mapping developed by the OGS during the 1980s were described. The first approach used basal till as a sample medium. This approach was developed between 1978 and 1982 in the KLIP (i.e., Kirkland Lake) area where there is deep, complex and continuous overburden. It was concluded that, although this approach was effective for geochemical mapping at the local scale, it was too impractical at the regional scale because of: 1) the complexity of Quaternary stratigraphy; 2) the unevenness of the bedrock surface; and 3) the difficulty (and high cost) of drill access in remote areas.

During the Southwestern Ontario Geochemical Survey program (1981 to 1983), successful regional geochemical maps of southwestern Ontario based on stream sediment were prepared. Important geochemical patterns in the survey data were described using composited data derived from 5 by 5 km or 10 by 10 km areas. This procedure produced strong and clear geochemical province and anomaly patterns for each of the 20 elements included in this survey. Most of these patterns could be explained in relation to the geology of the area and some by anthropogenic activity.

Since 1986, OGS geochemical mapping based on lake sediment core and water sampling has been sucessfully developed for use in areas of the Canadian Shield. This approach to regional geochemical mapping has proven to be a reliable, effective and economical component of the OGS mineral resource appraisal process.

Unfortunately, regional geochemical mapping based on stream or lake sediments cannot be completed in all of the areas where high mineral potential occurs in the province. For this reason, additional approaches to regional

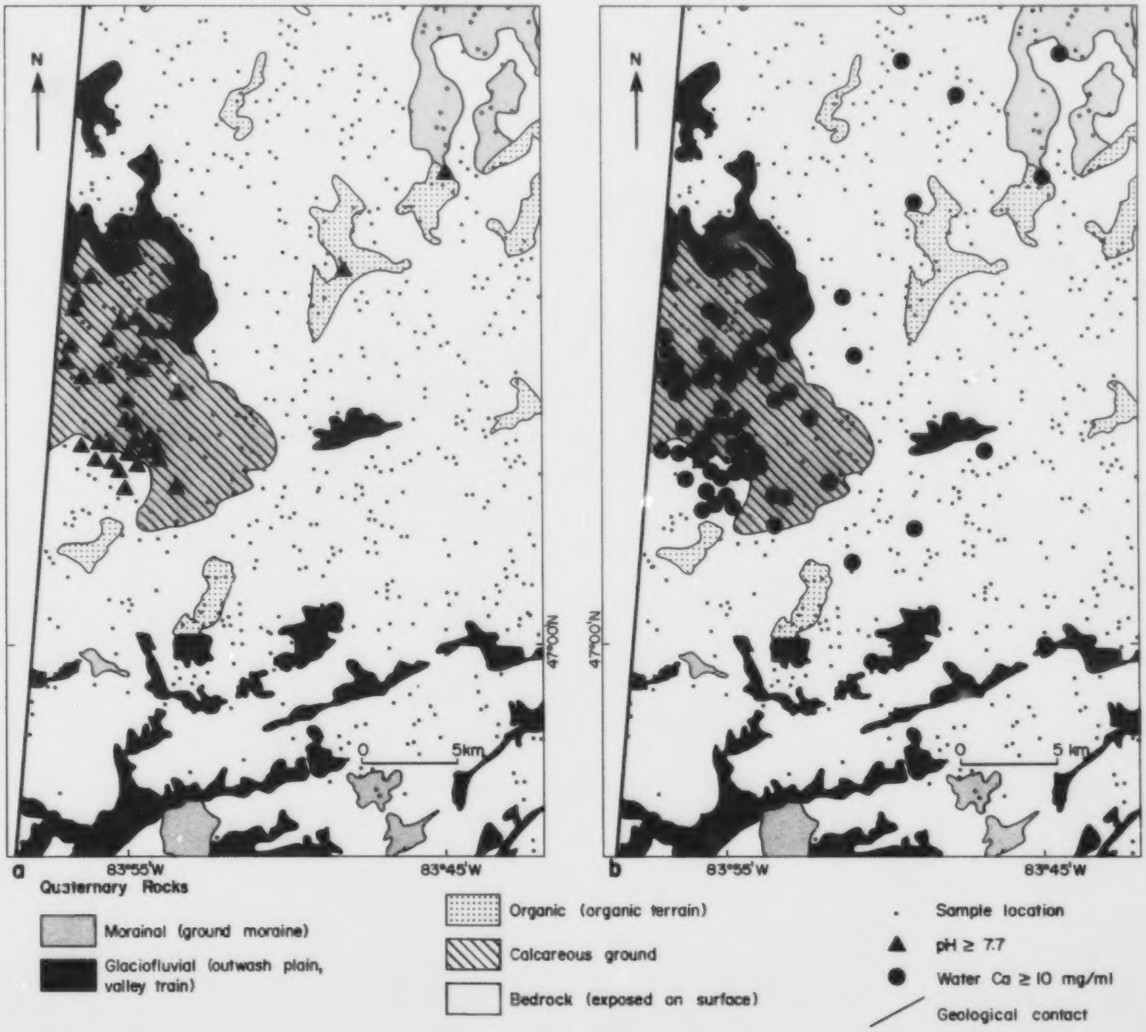

N
47°00'N
0 5km
a
83°55'W
83°45'W
N
47°00'N
0 5 km
b
83°55'W
83°45'W
Quaternary Rocks
Morainal (ground moraine)
Glaciofluvial (outwash plain, valley train)
Organic (organic terrain)
Calcareous ground
Bedrock (exposed on surface)
Sample location
pH ≥ 7.7
Water Ca ≥ 10 mg/ml
Geological contact

**Figure 27.21.** Quaternary geological maps of the Hanes Lake area showing: **a)** water pH values equal to or above 7.7, shown by large triangles; and **b)** water Ca values equal, or greater than 10 mg/ml, shown by large dots (*modified from* Fortescue and Vida 1990b).

geochemical mapping designed for 1) organic terrain, 2) areas of marl lakes and 3) areas of complex terrain (which need multimedia sampling), require development in the near future.

In general, the results of regional geochemical surveys presented in this chapter provide an introduction to the current state of the art in this type of mapping. It was not possible to include in this chapter examples of the presentation of regional geochemical map data using a GIS.

Part IV described 4 approaches to geochemical data verification methodologies and their viability. Some OGS environmental geochemistry research was then reviewed. Results show how it has aided in the development of modern regional geochemical mapping. This part ended with a discussion of the use of lake sediments in environmental geochemistry. Pre- and post-anthropogenic geochemical maps of the Herman Lake area (near Wawa) were chosen as an example of this trend. Because of the increased awareness of the importance of environmental geochemistry in Canada, this aspect of geochemical mapping is likely to increase in importance in the near future.

## Some Recent Developments in Geochemical Mapping

To conclude this chapter, it is appropriate to refer to a few of the new developments in geochemical mapping worldwide.

### THE GEOCHEMICAL SCOPE OF REGIONAL MAPPING

A total of 35 elements are currently included in OGS regional geochemical mapping (Fortescue and Vida 1989, 1990a, 1990b, 1991a, 1991b). Until recently, there was no international consensus among geochemists on the number of elements required for inclusion in geochemical maps. In 1990, this changed when Darnley (1990) and Xuejing (1990) stressed the importance of including 60 to 70 elements in global geochemical mapping. Many of the additional elements, for example, the Rare Earth Elements (REE), are now considered to be essential for mineral resource appraisal purposes, while other elements, such as iodine and selenium, are required for environmental geochemistry (*see* Table 27.6).

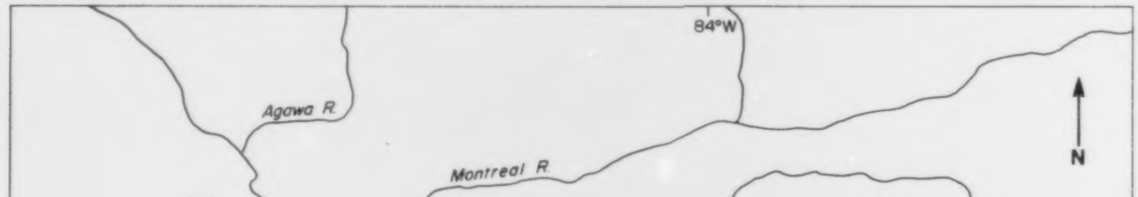

84°W
Agawa R.
Montreal R.
N
Agawa Bay
Lake Superior
Montreal River Harbour
Batchawana R.
Wlasy
Bracci
Butcher
Gapp
Gaudry
Hoffman
0
10
20 km
47°N
Lunkie
Nahwegezhic
Pine
Boundary of Hanes Lake study area
Hynes
Lamming
Menard
Township boundary
International boundary
Batchawana Bay
Canada
USA
Goulais R.
Ranger Lake
a

**Figure 27.22a. Map showing the location of the Hanes Lake study area. (*modified from* Fortescue and Vida 1990b).**

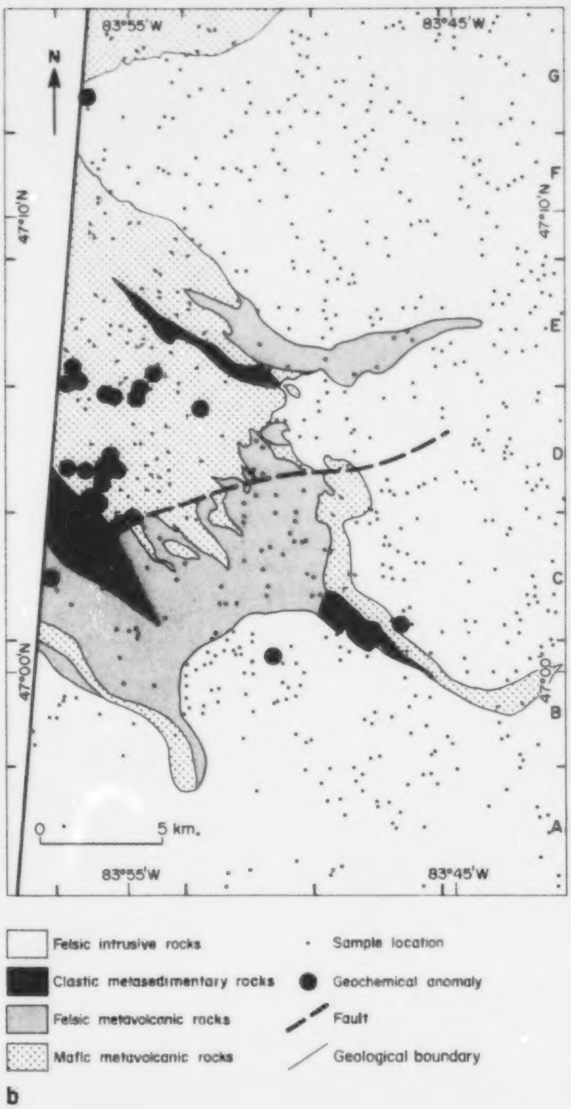

83°55'W
83°45'W
N
47°10'N
47°00'N
G
F
E
D
C
B
A
0
5 km.
Felsic intrusive rocks
Clastic metasedimentary rocks
Felsic metavolcanic rocks
Mafic metavolcanic rocks
Sample location
Geochemical anomaly
Fault
Geological boundary
b

**Figure 27.22b.** Map showing the location of geochemical sample sites and anomalies with respect to general geology of the Hanes Lake map area. Geochemical anomalies represent Clarke Index-I values greater than or equal to 2.0 Kk for arsenic (*modified from* Fortescue and Vida 1990b).

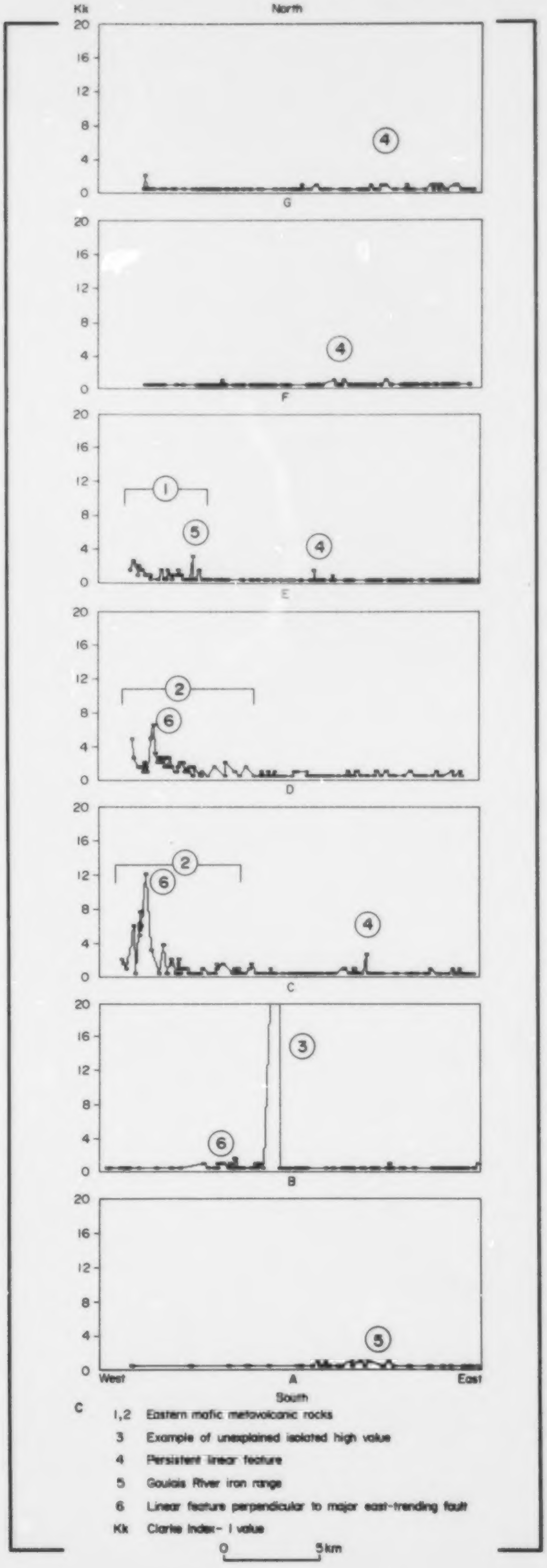


**Figure 27.22c.** Geochemical section diagrams showing a geochemical province pattern for arsenic in the Hanes Lake area. The sections are lettered to correspond with the appropriate area of the map in part b (*modified from* Fortescue and Vida 1990b).

**Table 27.4.** Geochemical data obtained from field verification of a pre–*Ambrosia* geochemical anomaly in the Trout Lake geochemical map area.

| Element | Order | $Datum_n$ | $Datum_k$ (GDS) | EG599 1988 | X1054 1989 | EG603 1988 | X1055 1989 | EG617 1988 | X1056 1989 |
|---|---|---|---|---|---|---|---|---|---|
| Mg | 1 | 3000 | 0.07 | 0.03 | 0.03 | 0.03 | 0.03 | 0.04 | 0.04 |
| Ti | 2 | 3000 | 0.11 | 0.03 | 0.05 | 0.05 | 0.05 | 0.05 | 0.06 |
| Mn | 3 | 3000 | 0.12 | 0.25 | 0.21 | 0.05 | 0.05 | 0.04 | 0.04 |
| Fe | 4 | 3000 | 0.12 | 0.19 | 0.15 | 0.06 | 0.07 | 0.07 | 0.08 |
| Na | 5 | 3000 | 0.13 | 0.04 | 0.03 | 0.04 | 0.04 | 0.07 | 0.05 |
| K | 6 | 3000 | 0.15 | 0.05 | 0.07 | 0.06 | 0.08 | 0.08 | 0.09 |
| Sr | 7 | 3000 | 0.15 | 0.05 | 0.05 | 0.04 | 0.05 | 0.07 | 0.07 |
| Ni | 8 | 3000 | 0.16 | 0.24 | 0.20 | 0.33 | 0.36 | 0.39 | 0.39 |
| Ca | 9 | 3000 | 0.18 | 0.19 | 0.21 | 0.18 | 0.21 | 0.17 | 0.17 |
| V | 10 | 3000 | 0.18 | 0.18 | 0.16 | 0.08 | 0.09 | 0.09 | 0.09 |
| Sc | 11 | 3000 | 0.20 | 0.16 | 0.15 | 0.16 | 0.15 | 0.24 | 0.20 |
| Co | 12 | 3000 | 0.21 | 0.31 | 0.24 | 0.28 | 0.24 | 0.28 | 0.24 |
| Cr | 13 | 3000 | 0.22 | 0.11 | 0.07 | 0.11 | 0.08 | 0.15 | 0.12 |
| Be | 14 | 3000 | 0.25 | 0.25 | 0.25 | 0.25 | 0.25 | 0.25 | 0.25 |
| Al | 15 | 3000 | 0.29 | 0.14 | 0.14 | 0.10 | 0.10 | 0.13 | 0.13 |
| Lu | 16 | 3000 | 0.37 | 0.19 | 0.19 | 0.19 | 0.19 | 0.37 | 0.37 |
| Th | 17 | 3000 | 0.37 | 0.25 | 0.20 | 0.25 | 0.23 | 0.25 | 0.36 |
| Ba | 18 | 3000 | 0.38 | 0.13 | 0.15 | 0.13 | 0.15 | 0.18 | 0.18 |
| Cu | 19 | 3000 | 0.49 | 1.85 | 2.01 | 1.88 | 2.07 | 3.76 | 4.13 |
| Au | 20 | 3000 | 0.50 | 0.25 | 0.25 | 0.25 | 0.25 | 0.25 | 0.25 |
| Pb | 21 | 3000 | 0.62 | 0.31 | 0.31 | 0.15 | 0.31 | 0.31 | 0.31 |
| La | 22 | 3000 | 0.69 | 0.38 | 0.20 | 0.29 | 0.17 | 0.55 | 0.35 |
| Hf | 23 | 3000 | 0.71 | 0.71 | 0.36 | 0.71 | 0.36 | 0.71 | 0.36 |
| P | 24 | 3000 | 0.72 | 1.16 | 1.13 | 0.35 | 0.42 | 0.29 | 0.28 |
| U | 25 | 2999 | 0.74 | 0.30 | 0.26 | 0.22 | 0.13 | 0.52 | 0.48 |
| Mo | 26 | 3000 | 0.83 | 0.83 | 0.83 | 0.83 | 0.83 | 1.67 | 0.83 |
| Zn | 27 | 3000 | 1.00 | 5.71 | 6.18 | 11.07 | 12.87 | 10.22 | 10.32 |
| Sb | 28 | 3000 | 1.00 | 1.00 | 1.00 | 1.00 | 0.50 | 1.00 | 0.50 |
| As | 29 | 3000 | 1.11 | 7.78 | 3.89 | 2.78 | 1.11 | 5.00 | 1.67 |
| Ta | 30 | 3000 | 1.18 | 1.18 | 0.59 | 1.18 | 0.59 | 1.18 | 0.59 |
| W | 31 | 3000 | 1.67 | 1.67 | 4.17 | 1.67 | 1.67 | 1.67 | 1.67 |
| Br | 32 | 3000 | 8.40 | 32.80 | 26.60 | 12.80 | 11.88 | 11.20 | 9.12 |

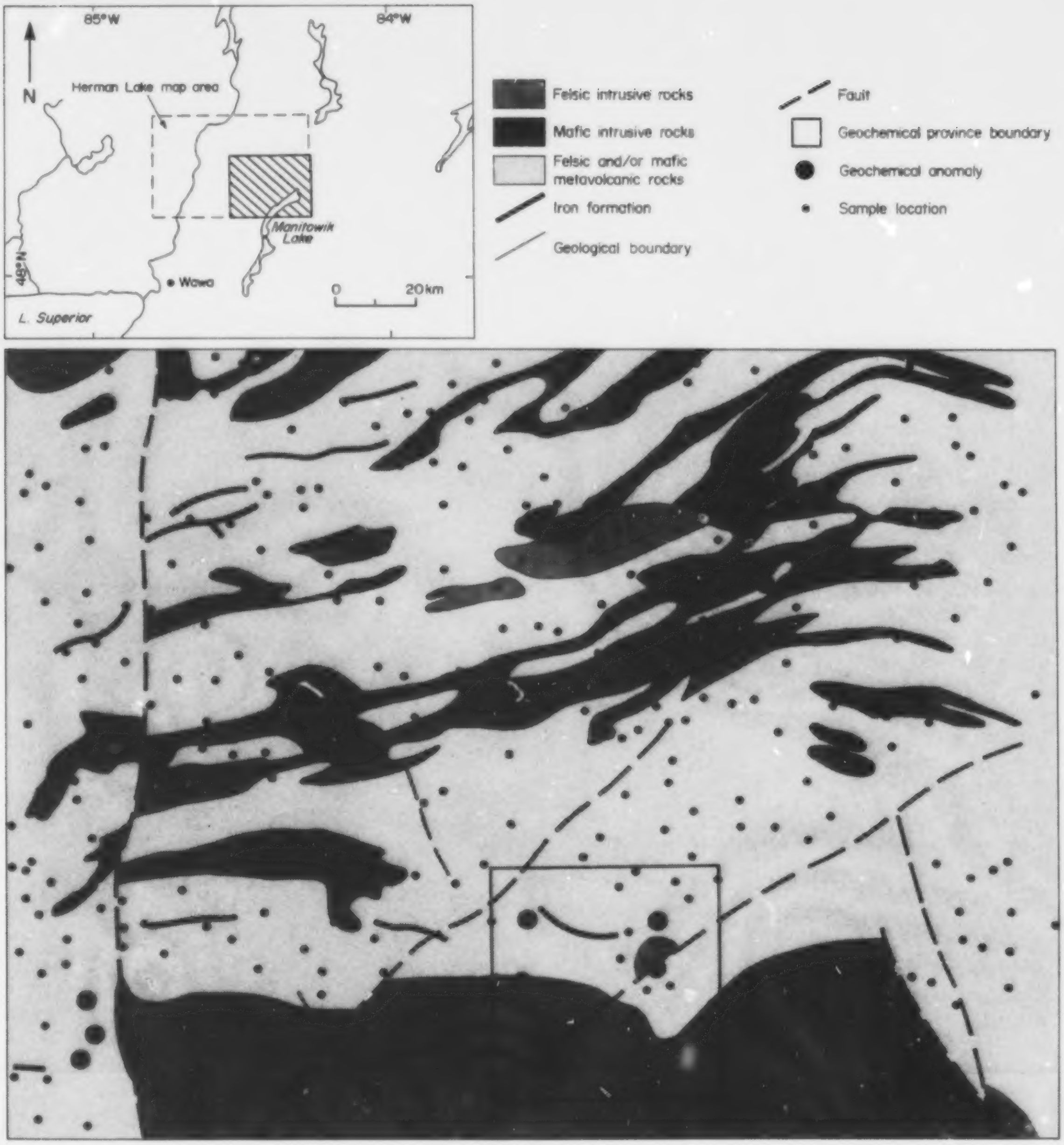

85°W
84°W
N
Herman Lake map area
Manitowik Lake
48°N
Wawa
0
20km
L. Superior
Felsic intrusive rocks
Mafic intrusive rocks
Felsic and/or mafic metavolcanic rocks
Iron formation
Geological boundary
Fault
Geochemical province boundary
Geochemical anomaly
Sample location

**Figure 27.23.** Map of part of the Herman Lake geochemical map area showing a geochemical province pattern for copper and the location of a pre-*Ambrosia* geochemical anomaly for copper which was verified by long-core drilling. Geochemical anomalies are represented by Clarke Index-I values greater than or equal to 2.0 for copper (*modified from* Fortescue and Vida 1990a).

Examples of organizations that are setting the new trend in geochemical mapping include: 1) the British Geological Survey, which listed 61 elements (excluding the REE) determined in groundwaters (Edmunds et al. 1989); 2) the Western European Geological Surveys (WEGS) planned for 70 elements to be routinely determined in 9000 overbank samples during a low density geochemical mapping project (Demetriades et al. 1990); and 3) the International Geochemical Mapping Project scientists who require a minimum of 65 elements in global-scale geochemical mapping (Darnley and Garrett 1990).

Elements included in future geochemical mapping are usually divided into 5 groups (*see* Table 27.6). If it is practical, the OGS geochemical maps prepared in the future should include all elements listed in groups 1 to 4 of the table.

**Table 27.5.** Geochemical data obtained from verification of a pre–*Ambrosia* geochemical anomaly by long core drilling of a sample site in the Herman Lake area (*modified from* Fortescue 1988b).

| Depth (inches) | LOI (%) | La (ppm) | Cu (ppm) | Mo (ppm) | Au (ppm) | As (ppm) |
|---|---|---|---|---|---|---|
| 6 | nss | 74 | 196 | 4 | <5 | 2 |
| 12 | 38.6 | 74 | 192 | 3 | 5 | 2 |
| 18 | 39.8 | 90 | 244 | 4 | <5 | 2 |
| 24 | 42.8 | 97 | 272 | 4 | <5 | 1 |
| 30 | 43.2 | 100 | 280 | 5 | <5 | 1 |
| 36 | 42.6 | 100 | 284 | 5 | <5 | 1 |
| 42 | 45.2 | 110 | 374 | 10 | <5 | 1 |
| 48 | 36.8 | 106 | 412 | 19 | <5 | <1 |
| 54 | 42.2 | 94 | 474 | 25 | <5 | 1 |
| 60 | 44.4 | 83 | 598 | 39 | 7 | 2 |
| 66 | 27.2 | 49 | 397 | 32 | <5 | 2 |
| 72 | 34.6 | 54 | 521 | 40 | <5 | 3 |
| 78 | 49.6 | 64 | 926 | 65 | 5 | 3 |
| 84 | 44.6 | 66 | 1335 | 46 | 11 | 5 |
| 90 | 55.0 | 58 | 1810 | 115 | 11 | 6 |
| 96* | 20.4 | 30 | 673 | 64 | 5 | 4 |
| 102 | 1.4 | 16 | 66 | 13 | <5 | 1 |
| 108 | 0.4 | 15 | 36 | 2 | <5 | 2 |
| 114 | 0.2 | 12 | 53 | 1 | <5 | 3 |
| 120 | 1.0 | 12 | 95 | 5 | <5 | 1 |

* Transition from organic to mineral matter.
LOI Loss on Ignition
nss Not sufficient sample

**Table 27.6.** Groups of elements included in regional geochemical mapping.

| | |
|---|---|
| 1. Major Elements (10): | Si, Al, Fe, Mg, Ca, Na, K, Mn, Ti and P |
| 2. Minor Elements (20): | Cu, Pb, Zn, Ga, Cr, Ni, Co, V, Sr, Ba, Sc, Y, La, Nb, (S), (Cl), Th, Rb, U and Zr |
| 3. Trace Elements (20): | As, Sb, Bi, Hg, Cd, Au, Ag, Se, Sn, W, Mo, B, F, Li, Be, Tl, In, Ge, Te and Ta |
| 4. Rare Earth Elements (14): | Ce, Pr, Nd, Pm, Sm, Eu, Gd, Tb, Dy, Ho, Er, Tm, Yb and Lu |
| 5. Platinum Group Elements (6): | Pt, Pd, Ir, Rh, Ru and Os |

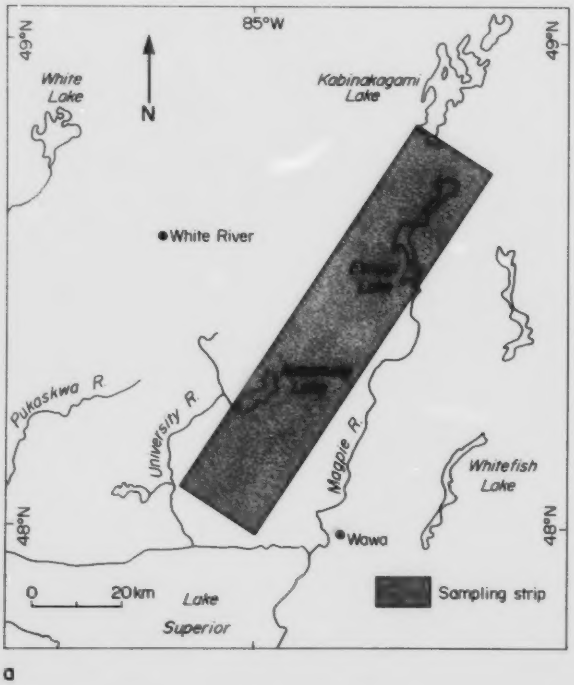

85°W
49°N
49°N
N
White Lake
Kabinakagami Lake
White River
Pukaskwa R.
University R.
Magpie R.
Whitefish Lake
48°N
48°N
Wawa
0
20km
Lake Superior
Sampling strip
a

**Figure 27.24a.** Map showing the location of the Wawa sampling strip (*modified from* Fortescue et al. 1981).

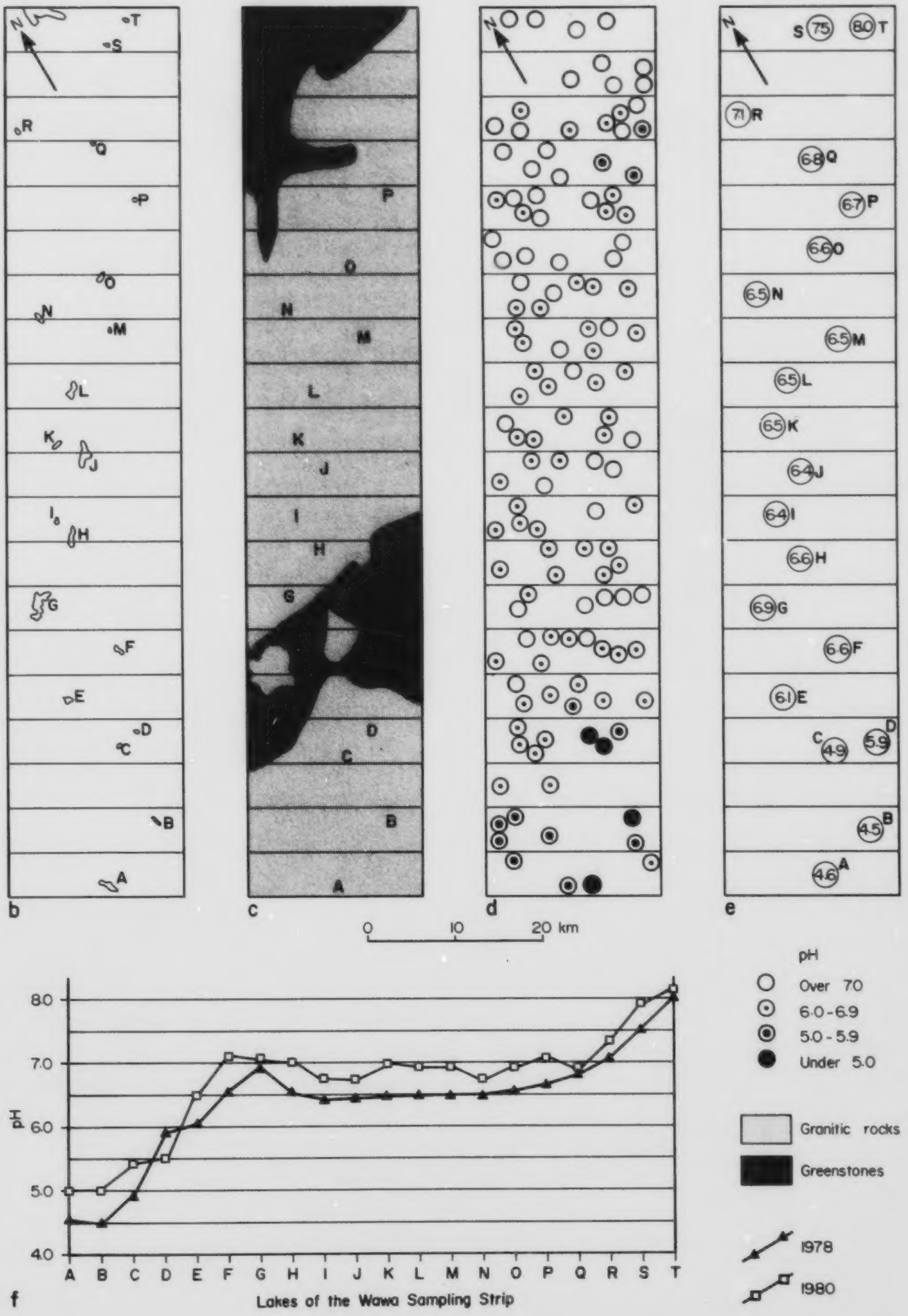

N
T
S
R
Q
P
O
N
M
L
K
J
I
H
G
F
E
D
C
B
A
b
c
d
e
S 7.5
8.0 T
7.1 R
6.8 Q
6.7 P
6.6 O
6.5 N
6.5 M
6.5 L
6.5 K
6.4 J
6.4 I
6.6 H
6.9 G
6.6 F
6.1 E
C 4.9
5.9 D
4.5 B
4.6 A
0 10 20 km
pH
○ Over 7.0
⊙ 6.0 - 6.9
◉ 5.0 - 5.9
● Under 5.0
Granitic rocks
Greenstones
1978
1980
pH
8.0
7.0
6.0
5.0
4.0
A B C D E F G H I J K L M N O P Q R S T
f
Lakes of the Wawa Sampling Strip

**Figure 27.24b to f.** Sketch maps of the Wawa sampling strip showing the: b) area of each study lake; c) bedrock geology; d) pH of all lakes tested in 1978; and e) pH of 20 lakes studied in 1978 by the Ontario Geological Survey. A separate graph f includes lakewater pH values recorded in 1978 and 1980 (*modified from* Fortescue et al. 1981).

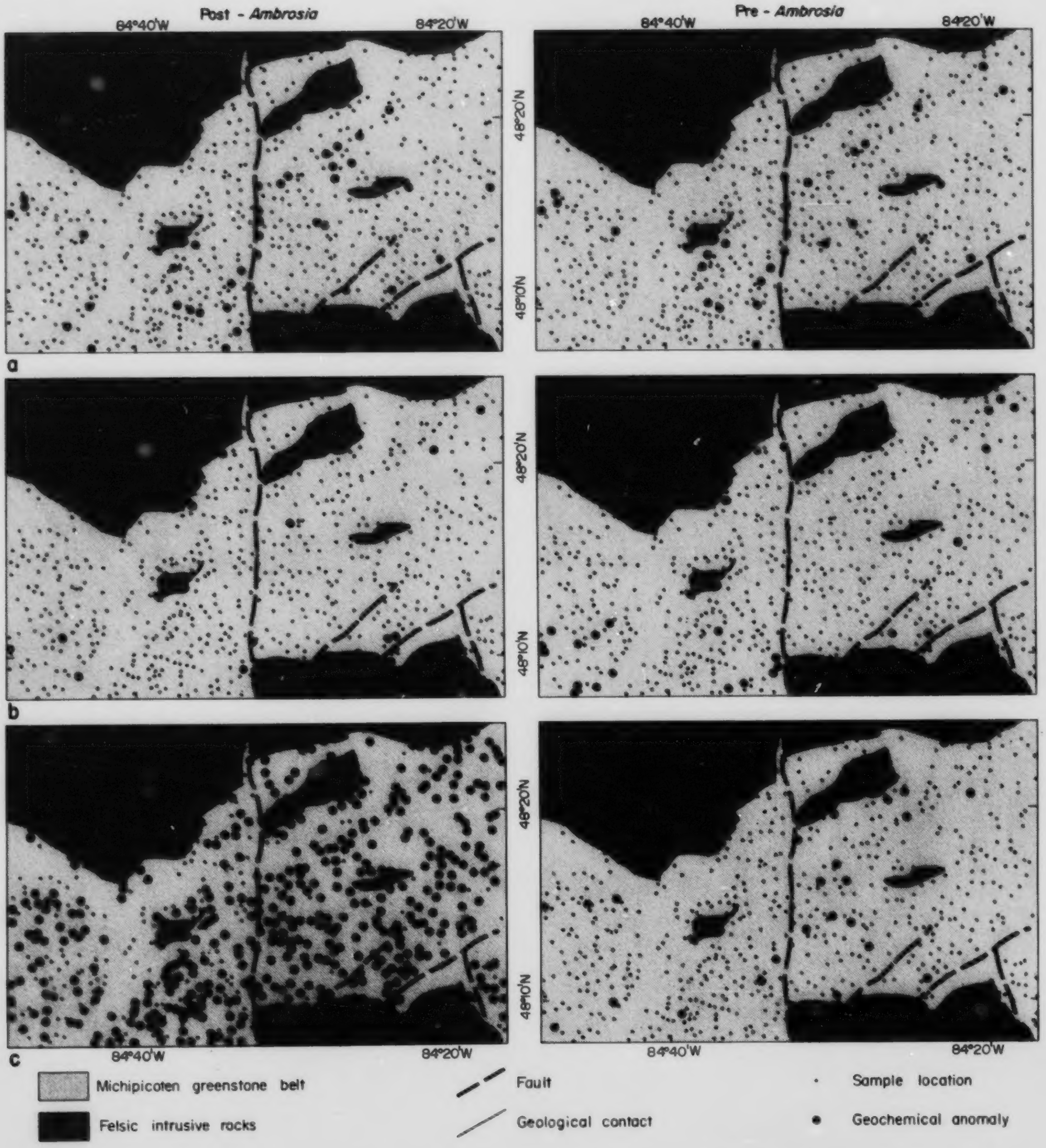

Post - *Ambrosia*
Pre - *Ambrosia*
84°40'W
84°20'W
48°20'N
48°10'N
a
b
c
Michipicoten greenstone belt
Felsic intrusive rocks
Fault
Geological contact
Sample location
Geochemical anomaly

**Figure 27.25a to c.** Pairs of generalized geological maps of the Herman Lake area showing post– and pre–*Ambrosia* geochemical province patterns. Geochemical anomalies are represented by Clarke Index-I values of: **a)** greater than or equal to 1.0 for manganese (Mn); **b)** greater than or equal to 2.0 for copper (Cu); and **c)** greater than or equal to 1.5 for lead (Pb). *See* Figure 27.17 for a location map of the Herman Lake study area (*modified from* Fortescue and Vida, in prep.).

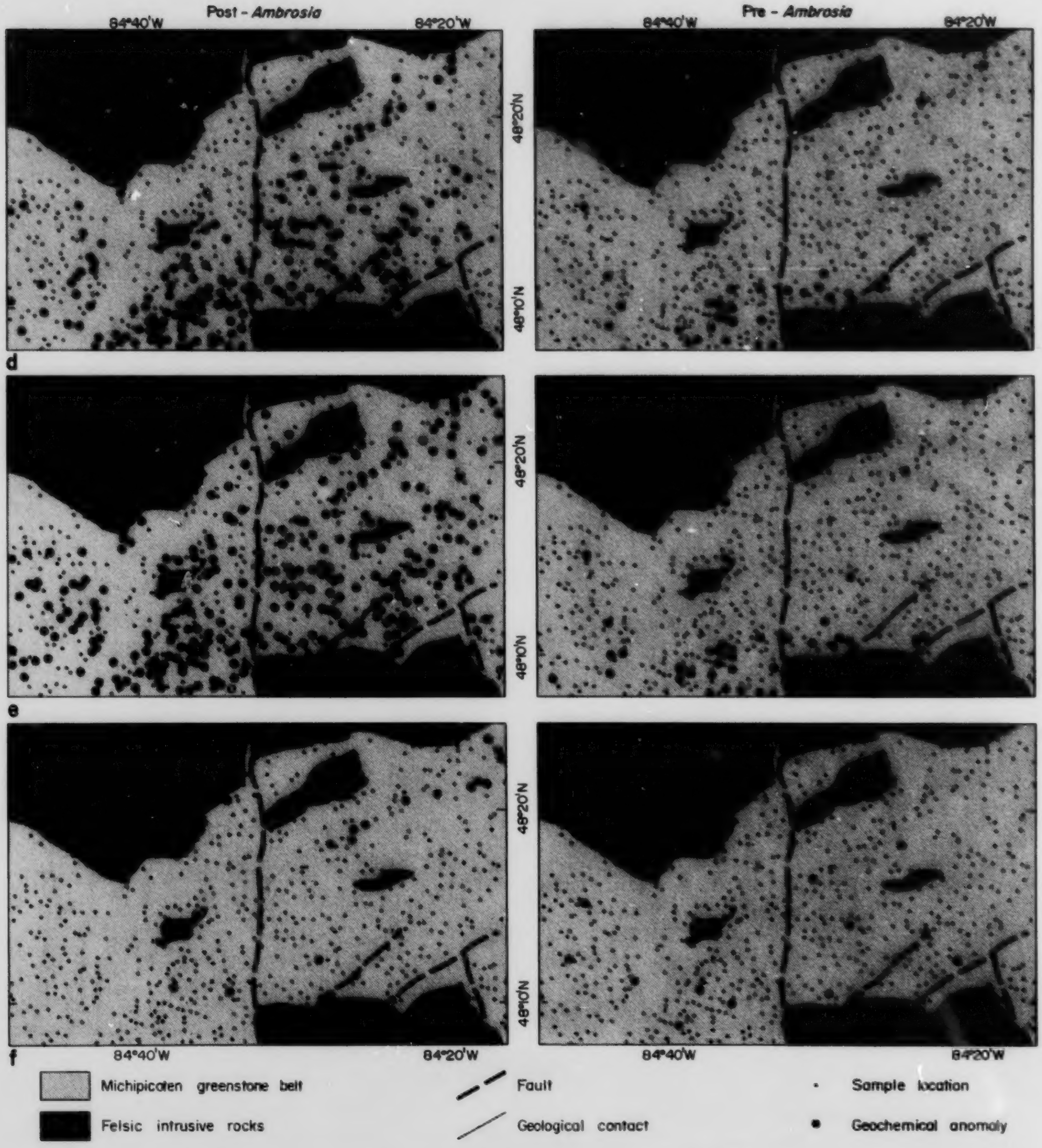

Post - *Ambrosia*
Pre - *Ambrosia*
84°40'W
84°20'W
48°20'N
48°10'N
d
e
f
Michipicoten greenstone belt
Felsic intrusive rocks
Fault
Geological contact
Sample location
Geochemical anomaly

**Figure 27.25d to f.** Pairs of generalized geological maps of the Herman Lake area showing post– and pre–*Ambrosia* geochemical province patterns. Geochemical anomalies are represented by Clarke Index-I values of: **d)** greater than or equal to 10.0 for arsenic (As); **e)** greater than or equal to 2.5 for antimony (Sb); and **f)** greater than or equal to 3.0 for gold (Au). *See* Figure 27.17 for a location map of the Herman Lake study area (*modified from* Fortescue and Vida, in prep.).

## THE DEDICATION OF A CHEMICAL LABORATORY TO OGS GEOCHEMICAL MAPPING

It is generally agreed among geochemists that, in order to obtain maps of high quality, the same standardized methodology must be rigorously applied throughout geochemical mapping. The standardization of a methodology of this kind for OGS regional geochemical mapping has already been described (*see* Figure 27.6).

It is clear that geochemical mapping programs in the 1990s will need to: 1) increase the number of elements determined in each sample; 2) increase the performance of chemical analysis for elements now included; and 3) reduce the time from sample collection to geochemical map and data base release.

The author proposed a novel solution to these problems based on a modern Mobile Laboratory Unit (MLU) to be constructed and dedicated entirely to the OGS geochemical mapping program (Fortescue 1990a). While parked at the base camp, the MLU would process 100 samples per day of lake sediment and lakewater (using a completely standardized methodology). It is considered feasible for the MLU to determine quantitatively some 50 elements in each lake sediment sample. In each lakewater sample, 20 elements would be determined quantitatively also.

At a field location, the MLU should be able to produce sufficient geochemical data for 2 regional maps without need for major resupply of gasses, chemicals and sample containers. During an environmental emergency, the MLU could produce geochemical maps within a few hours of sample collection, using an in-house geoscience GIS.

## THE GEOCHEMICAL MAP OF ONTARIO

In 1983, the author suggested that the OGS geochemical mapping program could lead to the production of a geochemical map of Ontario (Fortescue 1983b). At that time, a geochemical survey area of 890 000 km$^2$ was clearly impractical using a reconnaissance geochemical mapping methodology.

The recent development of low density-overbank geochemical mapping methodologies in Norway (Ottesen et al. 1989) and Finland (Koljonen et al. 1989) has demonstrated the feasibility of a modern geochemical map of Ontario. Such a map would require a minimum of 2000 sample sites. According to Ottesen et al. (1989), geochemical province patterns on low density geochemical survey maps are explained by: 1) the geochemical composition of bedrock; 2) the migration of elements in solutions and gasses of deep-seated origin, rising up through the lithosphere; 3) variations in the natural environment due to topography and climate; 4) long transported airborne elements of marine, volcanic and/or other natural derivation; and 5) local and/or regional contamination due to pollution from industry and other man-made sources. It is very likely that patterns of this kind would be identified on the proposed geochemical map of Ontario.

In August 1989, several overbank geochemical sample sites along the Goulais River (Searchmont area, north of Sault Ste. Marie) were discovered when B. Bolviken, the chief geochemist of the Norwegian Geological Survey, visited Ontario for a single day to search for sampling sites of this kind (Fortescue and Bolviken 1990). Samples were collected from one of these sites a few days later and geochemical data was subsequently obtained from them (Fortescue and Bolviken 1990). The feasibility of overbank sampling in the Sault Ste. Marie area was demonstrated by this experiment.

## Conclusions

1. The data obtained from regional geochemical mapping is playing an increasingly important role in the appraisal of Ontario's mineral resources.
2. The data obtained from regional geochemical mapping also plays an increasingly important role in environmental geochemistry in Ontario.
3. Due to variations in the landscape across the province, a single sample medium cannot be used to obtain a regional geochemical map of Ontario. Consequently, 3 different regional geochemical mapping methodologies were developed by the OGS during the 1980s. At least 3 more methodologies will be required to complete regional geochemical mapping of high mineral potential areas in the province.
4. In the 1980s, geochemical mapping evolved from the recognition of geochemical anomalies to the description of both geochemical provinces and geochemical anomalies. This change in emphasis has required: 1) a considerable increase in the scientific rigour of geochemical mapping methodologies; and 2) the introduction of advanced statistical techniques (and image processing and/or GIS) appropriate for the interpretation of multielement data geochemical sets.
5. Today, the OGS is in the vanguard of the development of geochemical mapping methodology. During the 1990s, this may be reflected by: 1) an increase in the number of elements included in regional geochemical maps; 2) the introduction of a Mobile Laboratory Unit dedicated to geochemical mapping; and 3) the production of a geochemical map of Ontario based on low density sampling.

## Acknowledgment

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## Chapter 28

# Characterization and Statistical Classification of Archean Volcanic Rocks of the Superior Province Using Major Element Geochemistry

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## Abstract

**Examination of the geochemical characteristics of Archean metavolcanic rocks is hampered by alteration and metamorphism which make geochemical classification of Archean metavolcanic rocks difficult. A review of various classification schemes based on chemistry has indicated that many are not suitable for Archean metavolcanic rocks.**

**The task of choosing a suitable geochemical classification scheme is based on the ability to separate altered from unaltered samples. Two classification procedures have been adopted in order to do this. Samples for the PETROCH data base of the Ontario Geological Survey were first subjected to classification procedures as outlined by Irvine and Baragar (1971) in order to pick up many of the effects of alteration. Alkali content, incompatible normative minerals and atypical abundances of major element oxides were used as the basis for rejecting altered samples. Atypical compositions were also identified by the use of Quantile-Quantile plots. Accepted subalkalic samples were then classified according to the technique outlined by Jensen (1976), which subdivides Archean volcanic rocks into komatiitic, tholeiitic and calc-alkalic magma suites. Komati-itic, tholeiitic and calc-alkalic reference groups were constructed from the "least altered" samples. Each magma clan was further subdivided for a total of 12 rock types, which represent the commonly accepted range of fractionated rock types within the 3 magma clans.**

**The statistical analysis of compositional data requires an approach which differs from conventional statistical analysis. The use of logratio transforms as suggested by Aitchison (1986) is a method for dealing with constant sum data. The use of logratio covariance structures allows for a valid interpretation of principal components analysis and subsequent canonical variate analysis and classification procedures.**

**The use of multivariate data analysis procedures provides an initial assessment of the chemical variation of the volcanic rocks in a reduced number of dimensions. This allows for a reduction in the number of variables required to describe the principal geochemical variation. Analysis of calc-alkalic, tholeiitic and komatiitic samples shows that each has a unique set of elements that make them relatively distinct from each other. Marginal overlap occurs between the 3 magma clans.**

**Canonical variate analysis was applied to the 3 reference magma clans and to the 12 rock types to test the statistical significance of the groups. The procedure indicates that all 3 magma clans and the 12 rock types are statistically distinct and can be used for allocating unknown samples for membership probabilities.**

**Allocation procedures assign probabilities of membership of an unknown sample into none, one or several of the reference groups. The index of typicality is based on the Mahalanobis distance of an unknown sample from each reference group centroid. If the Mahalanobis distance of a sample is within a specified limit that is determined by the dispersion (covariance) of the reference group, then the unknown sample will have a non-zero probability of group membership. If the Mahalanobis distance is greater than the dispersion of each reference group, then it will have a zero probability of membership to any of the groups. Alternatively, posterior probabilities assign group membership of an unknown sample based on a common covariance matrix for all of the reference groups. In this way, a sample will always be assigned a membership into one of the groups. Unknown samples were taken from the "altered" or rejected sample population and subjected to allocation procedures. Most of these samples do not belong to any of the reference groups due to the significant amounts of alteration that many have undergone. The atypical samples can be diagnostic of alteration zones around mineral deposits. Additional reference groups that define specific types of alteration are required in order to assign these unknowns.**

## INTRODUCTION

The classification of volcanic rocks has been the object of an International Union of Geological Sciences (IUGS) subcommittee since 1970 (Streckeisen 1979; LeMaitre 1984, 1989; Zanettin 1984; Le Bas et al. 1986) from which one of the accepted criteria for classification of volcanic rocks was the use of the Total Alkalis-Silica (TAS) diagram (Figure 28.1). Because of the high degree of mobility of sodium and potassium, the TAS diagram is one of the least suitable diagrams for classifying Archean volcanic rocks. This problem was acknowledged by the IUGS subcommission, but there were no recommendations made as to how to deal with samples that are altered or metamorphosed. This prompted the authors' investigation, and subsequently, a committee was established within the Ontario Geological Survey, based on the need to adequately classify and describe the extent of chemical variability of Archean volcanic rocks. The usefulness of an adequate Archean volcanic rock classification scheme has been discussed in Grunsky et al. (1986, 1987a, 1987b) and Easton et al. (1988). Various multivariate procedures were applied in order to characterize the chemical variation of the major Archean magmatic suites.

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The primary objective of the committee has been to define standard sets of representative Archean volcanic rocks that reflect distinct compositions in terms of magma genesis and tectonic environment. The ability to properly classify volcanic rocks and determine the nature of any associated alteration has significant implications for the inference of tectonic environment and mineral exploration strategies. Integral to this is the establishment of methods to determine which of the many discriminating diagrams available are of use in displaying the primary chemical variation of a volcanic sequence. In order to do this, the degree of alteration must be established by comparison with the "standards".

Almost all Archean volcanic rocks have undergone varying degrees of alteration and metamorphism. The mobility of the chemical constituents causes difficulty in using conventional discrimination diagrams for the purposes of assigning individual samples, or even groups of samples representing volcanic sequences, to the appropriate magma clan. Conventional methods of magma clan association utilize diagrams such as the AFM ($Al_2O_3$-$FeO^T$-MgO) diagram (Irvine and Baragar 1971), the AFTiM ($AlO_{1.5}$-FeO+$FeO_{1.5}$+$TiO_2$-MgO) molecular proportion plot of Jensen (1976) (Figure 28.2) and the ACM ($Al_2O_3$-CaO-MgO) diagram (Viljoen et al. 1982). Additional research into the variation of igneous systems and classification has been carried out by LeMaitre (1976c), Miyashiro (1974, 1978), Chayes (1966, 1979, 1980), Middlemost (1972) and Wilkinson (1986) to name but a few. During the 1970s and early 1980s, several researchers proposed diagrams that discriminated magma types and tectonic environments, as shown by Pearce and Cann (1971, 1973), Pearce and Norry (1979), Floyd and Winchester (1978), Winchester and Floyd (1975, 1976, 1977), Mullen (1983) and Pearce et al. (1975). Although many of the elements used are relatively immobile (Ti, Zr, Y, Nb, Sr, Sc, Ga), it has been demonstrated that in certain cases, the diagrams can fail to provide a proper classification (Prestvik 1982). There are even problems in drawing accurate variation diagrams as pointed out by Rickwood (1989).

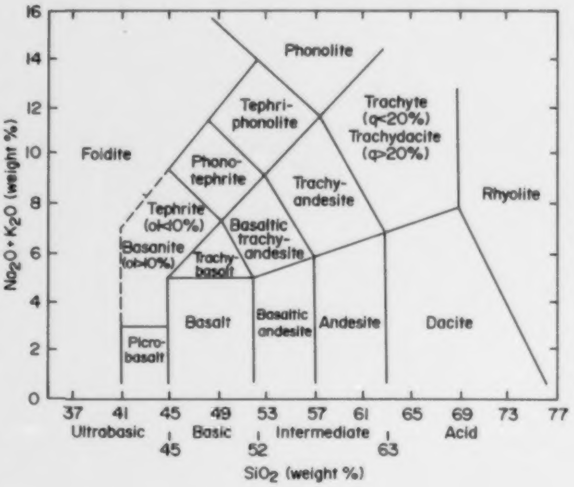
16
14
12
10
8
6
4
2
0
$Na_2O+K_2O$ (weight %)
Phonolite
Tephri-phonolite
Trachyte (q<20%)
Trachydacite (q>20%)
Foidite
Phono-tephrite
Trachy-andesite
Rhyolite
Tephrite (ol<10%)
Basaltic trachy-andesite
Basanite (ol>10%)
Trachy-basalt
Basalt
Basaltic andesite
Andesite
Dacite
Picro-basalt
37
41
45
49
53
57
61
65
69
73
77
Ultrabasic
Basic
Intermediate
Acid
45
52
63
$SiO_2$ (weight %)

**Figure 28.1.** Total Alkalis-Silica (TAS) diagram (*modified from* LeMaitre 1989) displaying the names of the rock types and $SiO_2$ saturation fields. q - quartz; ol - olivine.

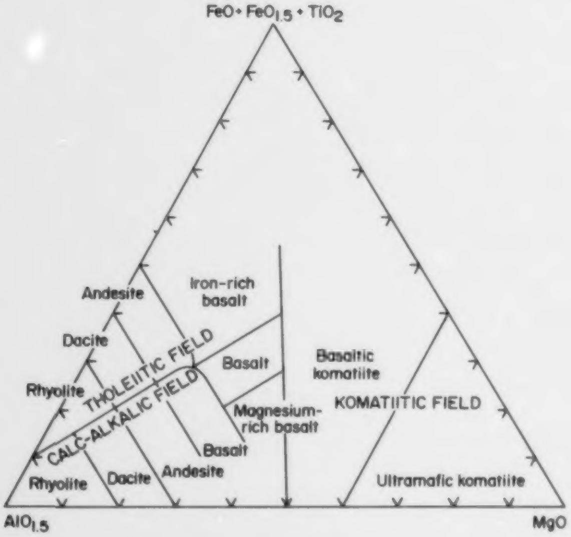
FeO+$FeO_{1.5}$+$TiO_2$
Andesite
Iron-rich basalt
Dacite
THOLEIITIC FIELD
Basalt
Basaltic komatiite
Rhyolite
CALC-ALKALIC FIELD
Magnesium-rich basalt
KOMATIITIC FIELD
Basalt
Rhyolite
Dacite
Andesite
Ultramafic komatiite
$AlO_{1.5}$
MgO

**Figure 28.2.** AFTiM molecular proportion diagram (*after* Jensen 1976) showing the magma clan and rock type boundaries.

For Archean volcanic rocks, many of these variation diagrams are not useful for classification due to the following reasons. First, these diagrams use only 2, 3 or 4 elements to display variation, and as a result, the variation of the other elements is completely ignored. Inferences based on variation of these selected elements can be misleading if other elements are varying. Secondly, the conventional diagrams such as the TAS ($SiO_2$ versus $Na_2O$+$K_2O$) or the AFM diagrams rely on elements (sodium, potassium and calcium) that are very mobile in environments of alteration and metamorphism. Thus, classification of Archean rocks using these methods must be scrutinized carefully and samples collected must represent the least altered or metamorphosed situation. In many Archean volcanic terrains this is impossible to achieve. Thirdly, classification methods that use trace elements for discriminating volcanic rocks in the oceanic environments of recent times do not seem to subdivide Archean volcanic sequences in any meaningful way. This may be due to the preserved Archean sequence representing "too compact" a tectonic record of the greenstone belt development, or, the Archean volcanic environment and subsequent geochemical environment was different from the patterns observed today.

Trace elements have been used as classifiers and discriminators in Archean sequences by Winchester and Floyd (1976, 1977), Floyd and Winchester (1978) and Gelinas et al. (1977, 1982, 1984). These approaches are effective at discriminating specific rock types and warrant investigation. One major drawback with the use of trace elements is that data bases composed of volcanic rocks seldom have consistent entries for trace elements, or when they do, the methods of analysis vary widely so that the differences between laboratory techniques can interfere significantly. Thus, the use of trace elements must be approached cautiously.

One of the primary purposes of this study is to provide acceptable limits of compositional variation for the 3 main magma clans of Archean volcanic rocks from the Superior Province of Ontario. Comparison of the chemistry of less understood sequences with well defined "least altered" reference groups can assist in characterizing the nature of volcanism and allow for an inference of the tectonic environment to be made.

This study is not intended to replace detailed petrographic work or provide a shortcut in determining the appropriate classification of an unknown sample. Rather, it is intended to be used in conjunction with suites of samples collected within an area in order to provide a basis of comparison with the least altered reference groups that have been established. Most importantly, this approach cannot replace detailed knowledge of field relationships that are critical in establishing relationships between petrochemical trends and tectonic environments that exist within geological domains. Gelinas et al. (1977) and M. Mellinger (Saskatchewan Research Council, personal communication, 1991) have pointed out that studies carried out in Quebec used petrographic criteria as an essential part of establishing the distinction between altered and least altered volcanic rocks.

The characterization and classification of volcanic rocks can be approached by 2 quantitative methods, namely, an investigation into the nature of the data by means of data analysis procedures, followed by a classification that is carried out using statistical procedures. Individual samples can be allocated membership into none, one or more than one of the reference groups using allocation procedures. The approach taken in this study is the utilization of both data analysis and statistical procedures for major element oxides only. Based on geochemical criteria and statistical procedures, least altered Archean volcanic rocks have been selected to compose the reference groups from which subsequent classifications are made.

## COMPOSITIONAL DATA, THE CLOSURE PROBLEM AND THE USE OF LOGRATIO TRANSFORMATIONS

A major obstacle in using and interpreting major element geochemistry is the closure problem. Compositional proportions (i.e., weight %) always sum to a constant (100%). Closed data, by definition, are points of the simplex, which is a part of n-dimensional real space (*see* Aitchison 1986, Section 2.2). The elements of points in the simplex are vectors of real numbers that satisfy the constant sum constraint.

The problem of closure has been acknowledged since the late nineteenth century (Pearson 1897), but has been largely ignored. Chayes (1960, 1966, 1970) and Chayes and Kruskal (1966) attempted to deal with the problem with limited success for specific applications. Only recently has Aitchison (1986) developed a theory and method for dealing with compositional data based on logratio transformations.

Standard numerical and statistical methods have been developed for data analysis where the values being considered are not closed. These methods cannot be applied to constant sum data without obtaining distorted results. In subcompositional analysis (examination of a subset of the elements that comprise a composition of a sample, i.e., AFM, AFTiM, TAS diagrams), these distortions can be significant (Aitchison 1986, p.33–36). Aitchison (1986) has developed a methodology for data analysis and statistical inference of compositional data by the use of logratio transformations. The use of these transformations projects the compositional data into the real number space, which allows standard statistical procedures to be applied. These methods are gaining popularity and an example of application to geochemical data is given by Butler and Woronow (1986) and Woronow and Butler (1986). More recent contributions to this approach are given by Aitchison (1990) and Woronow and Love (1990). The approach has also been extended into the theory of regionalized variables that is commonly used in ore reserve estimation (Pawlowsky 1989).

Additive logratio transformations are based on choosing one of the variables that comprise the major element oxide compositions and then dividing each major element oxide by this particular variable. This results in the loss of one of the variables, however, Aitchison (1986) has shown that the use of the additive logratio transformation does not result in the loss of any information about the compositional variation.

Given a data vector $x$ defined by $D$ major element oxides, a new transformed data vector $y$ can be calculated such that

$$y_i = \log(x_i/x_D), \quad \text{for } i = 1, \ldots, D-1$$

where $D$ = number of major element oxides.

Alternatively, the multiplicative logratio transformation can be used whereby each of the major element oxides is divided by the geometric mean of the elements comprising the sample. Given a data vector $x$ defined by $D$ major element oxides, a new transformed data vector $z$ can be calculated such that

$$z = \log(x_i/g(x)), \quad \text{for } i = 1, \ldots, D$$

where $D$ = number of major element oxides that form the constant sum, and

$g(x) = (x_1 \ldots x_D)^{1/D}$, the geometric mean of the data vector.

This transformation has the apparent advantage of maintaining all of the elements in the transformed data matrix. However, the associated correlation and covariance matrices of this transformation are singular and create difficulties when using statistical methods that require determinants or inverses of these matrices.

The analysis of subcompositions is an important aspect of geochemical studies. The problem of choosing suitable subcompositions that reflect significant compositional variation is also hampered by the problem of closure. The authors do not attempt to deal with this problem in this chapter. All of the major element oxides in the classification scheme will be considered. Aitchison (1986, Section 8.5) discussed the application of subcompositional analysis for determining the most significant variables. This chapter will address the classification of the major element oxide data using the additive and multiplicative logratio approach advocated by Aitchison (1986).

Other methods for dealing with compositional data have been proposed. Watson and Philip (1989) and Stanley (1990) preferred to treat closed data as a point lying on a hypershere. In the case of only 3 dimensions, the mathematics of data analysis and statistical inference are straight forward. For compositions of more than 3 variables, however, the mathematics becomes very involved, and at the time of writing, there were no public domain procedures or algorithms that allowed this approach to be used effectively.

Standard statistical methods of analysis have been developed on the basis that data is normally distributed. In geochemistry, this is not always the case. Elements such as titanium and potassium are commonly log-normally or non-normally distributed. This can cause difficulties in interpretation of statistical tests. Many techniques exist for transforming non-normal data into normal distributions. The most commonly used transformation is the log-normal transformation. However, as Howarth and Earle (1979) pointed out, the log-transformation may often be no better than no transformation at all. A more useful transformation is the Box-Cox family of power transformations (Box and Cox 1964). These transformations allow the fitting of a value of $\lambda$ for the equation:

$$z = (x^{\lambda} - 1)/\lambda \quad (\lambda \neq 0)$$

When $\lambda \rightarrow 0$, the transformation is replaced by

$$z = ln(x)$$

for all $x > 0$ where $x$ represents the abundance of the element in question.

One objection of applying transformations to multielement data is that each element may undergo a different transformation depending on the initial distribution. Thus, the original relationships of the data may be distorted. Howarth and Earle (1979) found that elements that have similar relationships prior to transformations tend to have similar $\lambda$ values, thus, maintaining the multielement relationships. However, this should be tested for each set of data where transformations are applied. As well, the ratios of data that have been subjected to power transformations may be difficult to interpret since they may not be normally distributed (Aitchison 1986, p.309). The use of power transformations has been investigated in this study. The results obtained in a canonical variate study of the reference groups are similar to the results obtained using the logratio transformations of Aitchison (1986). It is beyond the scope of this chapter to discuss the application and results of this procedure.

## DATA SELECTION

Our technique involves the assembly of a data base of major and trace element chemical data from Archean volcanic rocks of Ontario obtained from the PETROCH data base of the Ontario Geological Survey. The committee obtained 11 489 whole rock analyses from Archean volcanic rocks within the PETROCH data base. In addition, the Komatiite Database (Howe et al. 1987), composed of 255 whole rock analyses, was incorporated into the data base for a total of 11 744 analyses. The major element oxides and volatiles used are $SiO_2$, $Al_2O_3$, $Fe_2O_3$, FeO, MgO, CaO, $Na_2O$, $K_2O$, $TiO_2$, $P_2O_5$, MnO, $CO_2$, S, $H_2O^+$, $H_2O^-$ and LOI. Not all samples were analyzed for $CO_2$, S, $H_2O^+$ and $H_2O^-$, however, the total volatiles were measured as loss on ignition (LOI). LOI was used as the basis for assessing the volatile components in this study. All of the analyses in the PETROCH data base were carried out by the Geoscience Laboratories of the Ontario Geological Survey. As these samples accumulated over a period of over 20 years, the methods of analysis and precision and accuracy have changed. Thus, the sample population has built-in biases; a detailed study of the consistency of laboratory determinations over a twenty-year span is beyond the scope of this study. With the exception of $P_2O_5$, trends or patterns due to the methods and sensitivities of the analyses used, are masked by the major petrological trends and patterns.

Major element oxides with zero abundances were considered to be missing values in the PETROCH data base. Logratio transformations require all of the variables to be greater than zero, since the logarithm of zero does not exist. Aitchison (1986, Section 11.5) suggested several methods to cope with this problem; one approach is outlined in Appendix 1. The authors' experience with using replacement values indicated that samples which were assigned replacement values formed distinct populations separate from most of the data. This created difficulty in establishing homogeneous reference groups. Thus, all samples with major element oxides with zero abundances were eliminated. However, this creates problems for the komatiitic reference group, since it may not be unreasonable to encounter samples with $Na_2O$ or $K_2O$ abundances equal to zero. Eliminating komatiitic samples with zero abundances for $Na_2O$ and $K_2O$ results in inflated alkali values in the reference group.

Figures 28.3, 28.4 and 28.5 show 8667 samples (no oxides with zero abundances) selected from the PETROCH data base. The data has been contoured to assist in visualizing the density of sample distribution. The TAS diagram was gridded and contoured using an inverse square routine using cell sizes of 0.1 unit. The ternary diagrams were gridded and contoured as described by Smith (1968).

Figure 28.3 shows all of the PETROCH samples plotted onto the TAS diagram. Komatiitic rocks plot near the range of 40 to 45% $SiO_2$ and 0.0 to 0.5% $Na_2O$+$K_2O$. The tholeiitic basalts plot within the range of 45 to 55% $SiO_2$ and 1.0 to 4.0% $Na_2O$+$K_2O$. The fractionated tholeiitic rocks and the calc-alkalic rocks show a broad dispersion pattern of corresponding increases in $SiO_2$ and $Na_2O$+$K_2O$. The high degree of dispersion of the samples displays the degree of alkali enrichment in a large number of samples. The TAS diagram illustrates the large amount of dispersion due to alkali mobility in Archean volcanic rocks. With the exception of the obvious outliers, it is not apparent which samples are altered and which are close to an original, or least altered composition.

The AFM diagram (*see* Figure 28.4) has been conventionally used for displaying calc-alkalic trends and iron-enrichment trends in tholeiitic sequences (Irvine and Baragar 1971). The dividing line that separates the tholeiitic from calc-alkalic magma series is based on a set of Phanerozoic reference groups used by Irvine and Baragar (1971). This line may not be a suitable discriminator for Archean volcanic suites. Komatiitic rocks are not suited for the AFM diagram, as indicated in Figure 28.4, which shows all of the PETROCH samples. The diagram is useful for unaltered volcanic sequences, but typically in Archean volcanic rocks, sodium and potassium have been mobile, often making the diagram unreliable as a discriminator between magmatic suites. Thus, samples that plot into the calc-alkalic field may really be tholeiitic rocks in which there has been addition of $Na_2O$, $K_2O$ or both. This is particularly evident for the komatiitic samples. The data displays 3 broad trends. A tight cluster of ultramafic rocks are located along the MgO-$FeO^T$ edge of the diagram. The tholeiitic rocks form a large cloud of points that overlaps into the calc-alkalic field. The calc-alkalic rocks display a broad linear trend toward $Na_2O$ and $K_2O$ enrichment. Alkali enrichment due to fractionation cannot be distinguished from alkali metasomatism (*see* Figure 28.4).

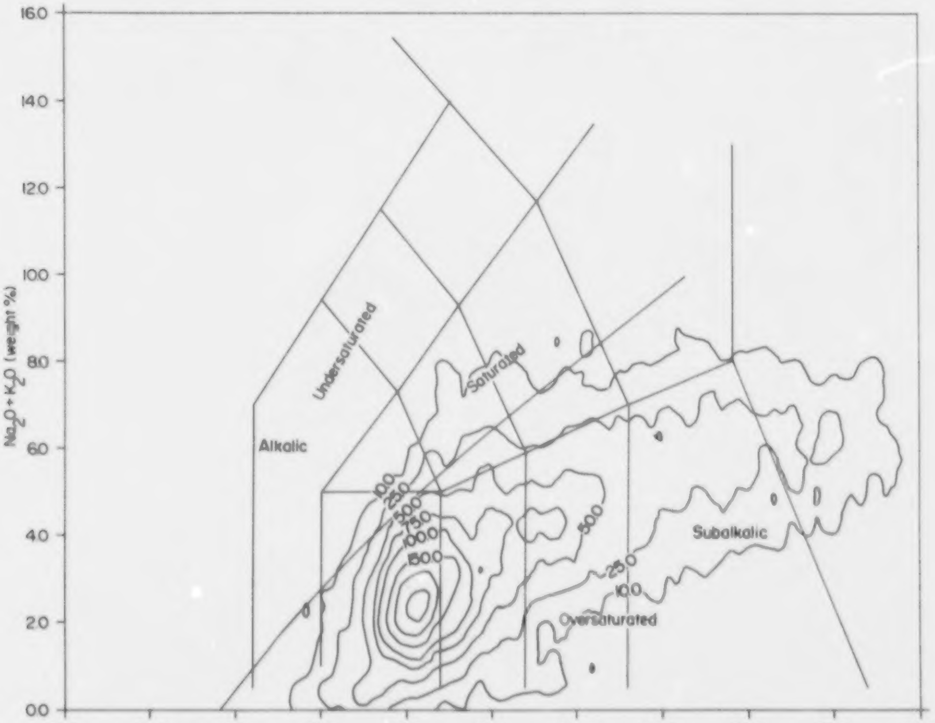

**Figure 28.3.** Total Alkalis-Silica (TAS) diagram (*modified from* LeMaitre 1989) for 8667 samples from the PETROCH data base. Note the large amount of dispersion and the large number of samples that plot within the alkaline part of the plot. The contour lines represent densities per unit cell in the figure. The figure was divided into 100 cells along the x-axis, 80 cells along the y-axis. Contour intervals are: 10, 25, 50, 75, 100, 150, 200 and 250 points per square unit.

As outlined above, the distinction between fractionated tholeiitic and calc-alkalic rocks cannot be easily resolved in an AFM diagram, as illustrated in Irvine and Baragar (1971, Figure 2). Thus, the decision to group a fractionated sample with the tholeiitic or calc-alkalic magma clan must be based on additional trace element analyses and on the associated rocks that occur within the sequence and cannot be based solely on the use of such a diagram.

The AFTiM molecular proportion diagram (*see* Figure 28.5) of Jensen (1976) is a useful diagram for displaying the komatiitic and basaltic komatiitic trends. It also displays the tholeiitic and calc-alkalic trends in a compressed form relative to the AFM diagram. The diagram is useful because it avoids the use of sodium and potassium, which are generally highly mobile elements in Archean volcanic sequences. This does not eliminate the problem of alkali enrichment, but rather, it hides the problem. The division boundaries of this diagram are based on Archean volcanic sequences of the Abitibi greenstone belt of Ontario. The division boundaries of calc-alkalic and tholeiitic basalts, andesites, dacites and rhyolites are somewhat arbitrary.

Samples that plot in the komatiitic region of the AFTiM diagram show a broad dispersion over the range of komatiitic compositions. Several of the samples display a relative increase in $FeO^T$ and $TiO_2$ and may represent suites of komatiitic rocks that are elevated in these elements relative to other suites. Recent work (Arndt and Jochum 1990) has shown that komatiitic compositions are commonly contaminated with other crustal material and due mainly to the high temperatures of these flows and the subsequent assimilation of surrounding country rock during eruption. This will undoubtedly create a high degree of variability in komatiitic compositions. Thus, komatiitic compositions must be interpreted cautiously.

Although the altered samples may appear to delineate magmatic trends in the AFM, AFTiM and TAS plots, there is no way of knowing that these samples actually belong to these magma clans. The classification of these samples has been based on the altered compositions; the original composition may have been from some other magma suite. Also, there is no way of judging just "how altered" a sample is. From a classification point of view, it may be desirable to

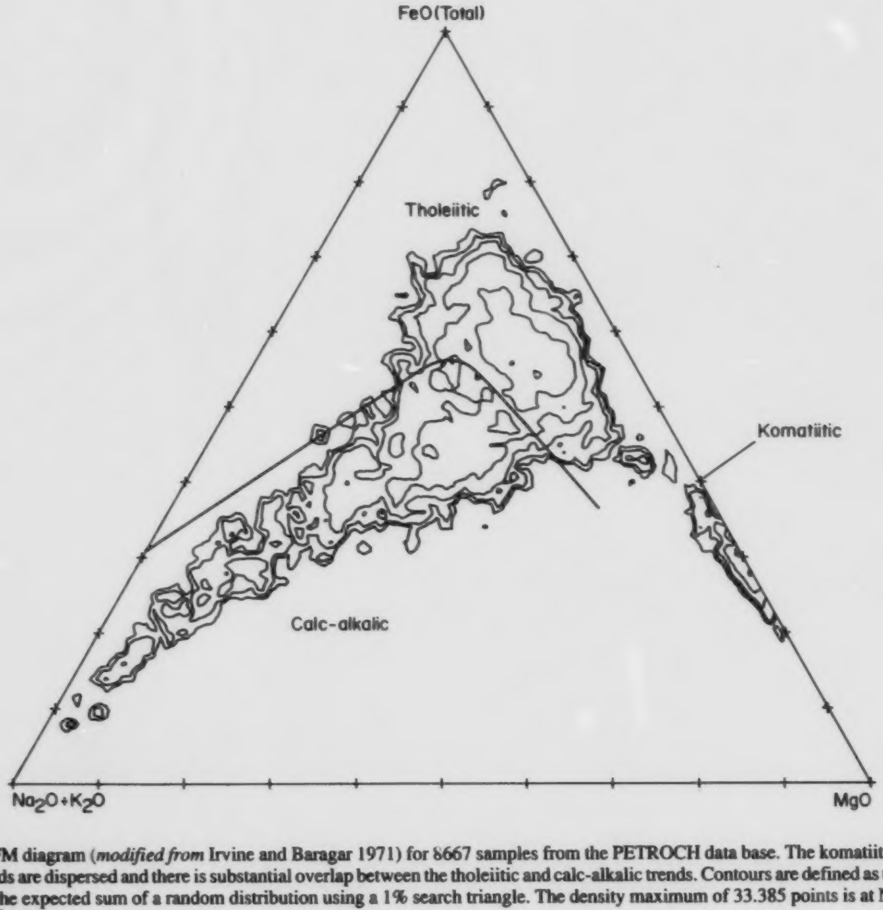

**Figure 28.4.** AFM diagram (*modified from* Irvine and Baragar 1971) for 8667 samples from the PETROCH data base. The komatiitic, tholeiitic and calc-alkalic trends are dispersed and there is substantial overlap between the tholeiitic and calc-alkalic trends. Contours are defined as units of standard deviation from the expected sum of a random distribution using a 1% search triangle. The density maximum of 33.385 points is at $Na_2O+K_2O = 12$ $MgO = 24$ $FeO^T = 54$.

know the degree of alteration so that any given sample can be accepted or rejected based on its degree of alteration.

## Selection of Reference Groups

Initially, the samples were classified using 2 standard geochemical classification schemes (Irvine and Baragar 1971; Jensen 1976). The criteria established by Irvine and Baragar (1971) makes use of normative mineral abundances. The classification of Jensen (1976) is based on the relative molecular proportions: $AlO_{1.5}$, $FeO_{1.5}$, FeO, $TiO_2$ and MgO. Both procedures assume a model based on an ordered sequence of crystallization.

The classification of Jensen (1976) was used to establish the magma clans because it is the more "robust" of the 2 methods. Aluminum, iron, titanium and magnesium are less mobile elements than many of the elements used to classify the various clans and rock types utilized by Irvine and Baragar (1971). The classification scheme only applies to subalkalic rocks. The other main advantage of the classification scheme of Jensen (1976) is that it recognizes iron-rich and magnesium-rich tholeiitic basalts, which are important rock types in Archean volcanic sequences (Basaltic Volcanism Project 1981, Section 1.2.1, p.12)

The choice of classification scheme for this study has been based on the collective experience and knowledge of the staff of the Ontario Geological Survey, in the volcanic sequences of the Superior Province of Ontario. The classification scheme of Jensen (1976) was developed from the examination of volcanic sequences in the Superior Province of Ontario and was thus considered appropriate for this study. The use of this diagram for areas outside of the Superior Province has not been studied closely. Baragar (Geological Survey of Canada, personal communication, 1991) has pointed out that high alumina basalts of the Circum-Pacific type plot in the intermediate calc-alkalic field of the AFTiM diagram. This suggests that its use for volcanic rocks other than Superior Province Archean greenstone belts may be questionable and should be used cautiously. Other methods of classification could have been chosen, such as the modification of the AFTiM diagram outlined by Viljoen et al. (1982); however, the characteristics of the South African greenstone sequences may be distinct from those of the Superior Province, and without a

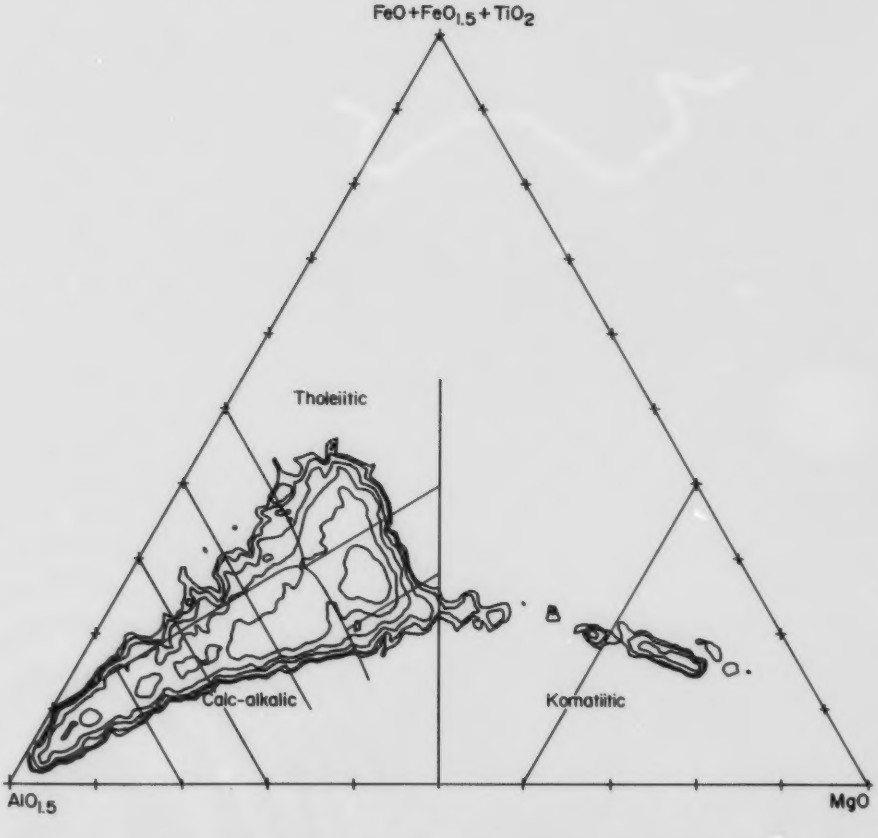

**Figure 28.5.** AFTiM molecular proportion diagram (*after* Jensen 1976) for 8667 samples from the PETROCH data base. Many samples indicate relative iron and titanium enrichment relative to aluminum and magnesium. The komatiitic, tholeiitic and calc-alkalic trends are dispersed and there is substantial overlap between the tholeiitic and calc-alkalic trends. Contours are defined as units of standard deviation from the expected sum of a random distribution using a 1% search triangle. Contour maximum represents 61.854 points at Al=45, Fe+Ti=27, Mg=28.

comparative analysis between the 2 areas, it would be unwise to adopt such a classification.

Because the AFTiM classification of Jensen (1976) is useful only for subalkalic volcanic rocks, alkalic rocks were rejected for this study. However, recent evidence suggests that calc-alkalic to alkalic sequences can occur within the same tectonic environment; it is possible that the 2 suites are genetically linked (Christiansen 1989; Colucci et al. 1989). Thus, the selection of choosing only subalkalic rocks may be misleading. This problem is further complicated because there has been a significant amount of alkali mobility in the Archean calc-alkalic sequences. Thus, it may be difficult to test this model by comparing samples that are classified as alkalic in Archean volcanic rocks. This hypothesis warrants further investigation. For the purposes of this study, the alkalic rocks were not considered any further.

## The Iron Correction Problem

In geochemically based classification schemes, the choice of iron ratios ($Fe^{+3}/Fe^{+2}$) poses a problem because the ratio affects the outcome of normative mineral calculations. In most Archean volcanic rocks, this ratio has changed through alteration and metamorphism. Irvine and Baragar (1971) suggested that iron correction be based on the amount of $TiO_2$ and defined the following formula:

if ($FeO>(1.5+TiO_2)$) then

$Fe_2O_3=Fe_2O_3-(Fe_2O_3-(TiO_2+1.5))$

$FeO=FeO+(Fe_2O_3-(TiO_2+1.5))*0.8998$

LeMaitre (1976b) used modern volcanic sequences to adjust iron ratios by a regression of these ratios on the alkalis. Such a correction would be unreasonable to apply to Archean rocks because of the mobility of the alkalis. More recently, Middlemost (1989) published a diagram of $Fe^{+3}/Fe^{+2}$ ratios for modern volcanic rocks based on the TAS diagram. The authors have adapted these ratios to the classification of Jensen (1976). They are listed in Table 28.1.

To use the iron ratios of Middlemost (1989), an initial estimate is required to determine the rock type and hence the use of appropriate iron ratio. Thus, a two-step process is involved and the authors' classification scheme involves the following calculations.

### STEP 1

1. Eliminate data where $SiO_2$, $Al_2O_3$, $Fe_2O_3$, FeO, MgO, CaO, $Na_2O$, $K_2O$, $TiO_2$, $P_2O_5$ and MnO were equal to zero. A total of 3077 samples were eliminated in this step, resulting in 8667 samples. Initially, values of zero for $P_2O_5$ and MnO were considered to represent values less than detection limit and were assigned replacement values as described in Appendix 1. Subsequent analysis of samples with these replacement values created difficulty in the interpretation of the data (*see* Principal Components Analysis Applied to the Geochemistry of the Reference Groups) and resulted in deleting all samples with zero abundances for $P_2O_5$ and MnO.

**Table 28.1.** Iron ratios used for classification.

| Rock Type | Code | Iron Ratio ($Fe_2O_3/FeO$) |
|---|---|---|
| Ultramafic komatiite | KOUM | 0.15 |
| Basaltic komatiite | KOBA | 0.15 |
| Fe-rich tholeiite | THFE | 0.20 |
| Mg-rich tholeiite | THMG | 0.20 |
| Tholeiitic basalt | THBA | 0.50 |
| Tholeiitic andesite | THAN | 0.35 |
| Tholeiitic dacite | THDA | 0.40 |
| Tholeiitic rhyolite | THRH | 0.50 |
| Calc-alkalic basalt | CABA | 0.20 |
| Calc-alkalic andesite | CAAN | 0.20 |
| Calc-alkalic dacite | CADA | 0.35 |
| Calc-alkalic rhyolite | CARH | 0.40 |

2. Recalculate weight % oxide data into adjusted and cation percentages after applying the $Fe^{+3}/Fe^{+2}$ ratio correction outlined by Irvine and Baragar (1971).
3. Compute the Barth-Niggli normative mineral calculation.
4. Classify the samples using the method of Irvine and Baragar (1971). The alkalic or subalkalic nature of the sample is determined.

### STEP 2

1. Apply the $Fe^{+3}/Fe^{+2}$ corrections as proposed by Middlemost (1989) to the analysis and recompute according to the classification of Irvine and Baragar (1971). If the sample is not subalkalic, then it is rejected.
2. Apply the classification scheme of Jensen (1976).

## Reference Group Selection Using Chemistry and Normative Minerals

Selection of the least altered samples for reference groups was based on the criteria previously outlined by other authors (Gelinas et al. 1977). The classification scheme of Irvine and Baragar (1971) was used to distinguish peralkalic, alkalic and subalkalic rocks. From an initial group of 8667 samples, the following criteria were used for selecting and/or rejecting samples.

Peralkalic Rocks: Samples of a peralkalic character as defined by the classification scheme of Irvine and Baragar (1971) were rejected (3 samples).

Alkalic Rocks: Samples that plot in the alkalic fields of the TAS and Ol′-Ne′-Q′ diagrams of Irvine and Baragar (1971) were rejected. This included 922 samples that were classified as alkalic (398 samples) and potassium-rich alkalic rocks (152 samples). In addition, a number of samples were classified as alkalic and contained the following atypical normative minerals: calcium silicate (22 samples), hypersthene (233 samples), kaliophilite (27 samples), sodium silicate (10 samples) and quartz (69 samples). A total of 11 samples were classified as alkalic, but failed to classify properly and were labelled as "unassigned".

Subalkalic Rocks: Many of the samples that were classified as subalkalic were rejected because of the presence of normative minerals that were not considered to be compatible for these rocks. These minerals included: hematite (134 samples), sodium silicate (1 sample), wollastonite (48 samples) and nepheline (1 sample). Some samples, although classified as subalkalic, could not be assigned to any particular rock type and were classified as unassigned (7 samples).

Potassic samples (potassium rich) as determined from Irvine and Baragar (1971, Figure 8) were considered to be atypical of Archean volcanic rocks and were rejected (1214 samples).

Samples that contained normative corundum were considered to be altered. Initially, all samples with normative corundum were rejected; however, it was found that a disproportionate number of fractionated calc-alkalic rocks were rejected. Many of these samples were found to have abnormally low CaO. This suggests that CaO has been stripped from many of these rocks. This resulted in rejection of 1870 samples. The procedure was subsequently modified to accept calc-alkalic rocks with normative corundum abundances of less than 1.0%. Samples with normative corundum of less than 1.0% were subsequently included in the accepted calc-alkalic groups. Many of these samples were subsequently rejected in additional reference group selection procedures (Quantile-Quantile [Q-Q] plots).

Unassigned Rocks: Some samples were of such atypical character that they could not be classified using the methods of Irvine and Baragar (1971), or the method of Jensen (1976). These samples were set as being unassigned and were also rejected (7 samples).

Using these selection procedures, the number of least altered samples was reduced to 4460, distributed between the 3 main magma clans and rock types as shown in column A of Table 28.2. A total of 4207 samples were classified as altered.

The classification schemes are based on the assumption that the major element oxide abundances represent the mineralogy that uniquely defines a volcanic rock. Classification schemes are based on a sequence of crystallization; there is an assumed equilibrium between the chemistry and the corresponding mineralogy. As well, the stoichiometry of the model forces the elements to combine in specific forms. Although AFM, AFTiM and TAS diagrams use the major element oxide chemistry, the stoichiometry of an assumed fractionation model is implied in the boundary lines that are drawn. However, many of the samples in the PETROCH data base have been rejected because the observed geochemistry departs from the abundances required to form the necessary normative minerals, and yet many of these samples fall within the accepted boundaries of the diagrams. This also implies that each major element oxide will have a range of values, based on normative mineral calculations, in which the sample will be accepted as being classified successfully. If the element abundance falls outside of this range of values, then it is unlikely to represent the rock type that is indicated in the diagrams.

Examples of this are as follows. Consider a rock that has undergone extensive carbonatization. In many volcanic environments, calcite replaces calcic plagioclase, but despite the extensive recrystallization, the relative abundances of aluminum, iron, magnesium, sodium and potassium can remain unchanged. Although the rock has undergone significant alteration, it will still be considered an acceptable sample when classified using the procedures outlined by Irvine and Baragar (1971), the TAS diagram or the AFTiM diagram. This sample cannot be considered as representing a least altered analysis for the reference group, although it will plot within the accepted boundaries on the commonly used discrimination diagrams.

Elements that are not directly involved in the classification will also affect the results as the data is closed. Thus, if a rock has greater than 5% $K_2O$ and it is a basaltic rock, it would be considered unacceptable as a sample to represent a reference group of tholeiitic basalts. Depending on the classification scheme used, the elevated $K_2O$ abundance may not be detected and yet, this unacceptable sample would be included in one of the reference groups.

The distinction between the fractionated tholeiitic and calc-alkalic rocks must be cautiously applied. In many cases, these fractionated rocks have similar characteristics. It is not wise to base the classification solely on the position within the AFM or AFTiM diagrams. The association of the samples within the sequence of volcanic rocks is of greater significance. If a rhyolitic sample is found within a tholeiitic sequence, it is more likely that it is part of a fractionating tholeiitic magma, even though it may plot within the calc-alkalic rhyolitic field on the AFTiM diagram. This has been

**Table 28.2.** Tally of initial and final least altered suites of samples.

| Code | Rock Type | Code | A<br>No. of Samples | B<br>No. of Samples |
|---|---|---|---|---|
| 1 | Ultramafic komatiite | KOUM | 280 | 83 |
| 2 | Basaltic komatiite | KOBA | 339 | 72 |
| 3 | Fe-rich tholeiite | THFE | 715 | 258 |
| 4 | Mg-rich tholeiite | THMG | 177 | 74 |
| 5 | Tholeiitic basalt | THBA | 1227 | 683 |
| 10 | Tholeiitic andesite | THAN | 171 | 49 |
| 11 | Tholeiitic dacite | THDA | 96 | 25 |
| 12 | Tholeiitic rhyolite | THRH | 36 | 23 |
| 6 | Calc-alkalic basalt | CABA | 711 | 265 |
| 7 | Calc-alkalic andesite | CAAN | 376 | 138 |
| 8 | Calc-alkalic dacite | CADA | 211 | 113 |
| 9 | Calc-alkalic rhyolite | CARH | 121 | 62 |
| | Total | | 4460 | 1845 |
| | Totals by magma clan | | | |
| | Komatiites | KOM | 619 | 155 |
| | Tholeiites | THOL | 2422 | 1112 |
| | Calc-alkalic rocks | CALC | 1419 | 578 |
| | Total | | 4460 | 1845 |

A - Prior to Q-Q and $\chi^2$ plot trimming
B - After Q-Q and $\chi^2$ plot trimming

KOM = KOUM + KOBA
THOL = THFE + THMG + THBA + THAN + THDA + THRH
CALC = CABA + CAAN + CADA + CARH

a problem in the selection of samples representing the reference groups for this study. Our selection process has been based solely on the automatic classifications outlined above. No consideration has been given to the stratigraphy of the samples included in the reference groups.

## Reference Group Refinement Using Q-Q Plots

The range and distribution of major element oxide abundances can be investigated by the use of Quantile-Quantile (Q-Q) plots for each of the major element oxides (marginal distributions) in the rock type and magma clan reference groups. Q-Q plots (ranked observations versus quantile values for a normal distribution) provide a diagrammatic description of the sample population. Ideally, samples that are normally distributed, when compared to a theoretical normal distribution, plot as a straight line. The main advantage of the Q-Q plot is that each individual sample is plotted and thus the detailed characteristics of groups of samples that are of interest can be observed (cf., Figure 28.6).

Samples that are not part of the same population (outliers) are markedly different and appear as individuals or groups of samples with a slope or trend that is different from the main population. Samples that do not form part of the continuous population are deleted from the reference group. It is assumed that the major element oxides that define the geochemical compositions are continuous in their frequency distributions.

An important consideration is the case where an element, such as $K_2O$, forms part of a continuous distribution with values that are known to be unacceptably high or low. In this case, some arbitrary decisions must be made as to what constitutes an "outlier". Such distributions will be truncated and result in poorer estimates of the parameters of the distribution.

When sample populations display a curved trend, the application of power transformations can be used to shift the frequency distribution to normality. Transformations are often necessary for element oxides such as $K_2O$, $Na_2O$ and the volatile components. Usually, a logarithmic transformation results in a normal frequency distribution. Transformations were applied according to the method described by Howarth and Earle (1979).

An important assumption in the use of these methods is that the sample populations are multivariate-normally distributed. Even if the marginal distributions (distribution of each major element oxide) are normal, this is no guarantee that the multivariate population is normal. However, if the marginal distributions have outliers and are highly skewed, the multivariate population will be certain to have even more non-normal characteristics. Thus, an important criterion in applying statistical procedures is the elimination of outliers and an approximation to a normal distribution.

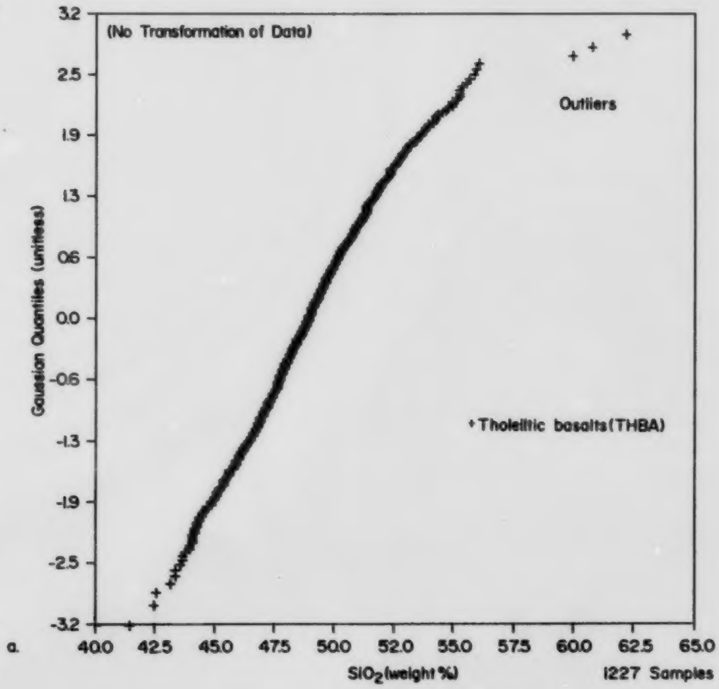

**Figure 28.6.** a) Quantile-Quantile (Q-Q) plot of $SiO_2$ for 1227 tholeiitic basalts (THBA) showing the outliers. Notice how the presence of outliers distorts the shape of the trend of the data. b) Q-Q plot of $SiO_2$ for 683 tholeiitic basalts (THBA) with the outliers removed. The trend of the data follows a straight line without the outliers.

Figure 28.6a shows $SiO_2$ for 1227 tholeiitic basalts (THBA) displayed on a Q-Q plot. These plots were obtained from procedures developed by Campbell (1986). The ordinate represents the ordered values of $SiO_2$; the abscissa represents the quantile values of a normal distribution. If the values of $SiO_2$ are normally distributed, then the samples plot as a straight line with a slope of 1. In this case, the line is reasonably straight, but 3 individual samples appear as outliers away from the main distribution of points. These samples are not part of the main trend of the data and were eliminated from the reference group.

Figure 28.6b shows the Q-Q plot of $SiO_2$ for 683 tholeiitic basalts with the outliers removed. The samples plot on a straighter line, and with a slope closer to a value of 1. As this sample closely resembles a normal population, standard statistical procedures can be applied to this population with meaningful results. However, note that the population has been reduced by almost 50%. This is due to the fact that outliers were also eliminated from Q-Q plots for the other major element oxides.

Q-Q plots were first inspected for the primary element oxides, FeO, MgO, $Al_2O_3$, CaO, $Na_2O$, $K_2O$, $SiO_2$ and LOI. The oxides $Fe_2O_3$, $P_2O_5$ and MnO were considered as secondary distributions. The primary elements were examined first, followed by examination of the secondary distributions. In most cases, the frequency distributions formed continuous lines typical of the primary frequency distributions. The main objective of outlier elimination was to establish continuous unimodal distributions for FeO, MgO, $Al_2O_3$, CaO, $SiO_2$, $Na_2O$ and $K_2O$ by trimming outliers from both ends of the curves.

Gelinas et al. (1977) imposed a limit of 4.0 weight % on the total volatiles ($CO_2$+$H_2O$) permissible in an analysis. The use of Q-Q plots indicates that in some cases this constraint is unrealistic and truncates the distribution of LOI. Q-Q plots of LOI for the basaltic rocks consistently defined normal populations (after transformation) with mean values in the range of 5 to 7 weight %. After trimming, the mean volatile abundances remained above 4%, which the authors consider acceptable. This is partly due to the fact that rocks that have not been extensively altered, in which the addition of $CO_2$ or $H_2O$ results in the development of amphibole or carbonate, may still have a balanced stoichiometry of the original cations and thus be an acceptable representative of the least altered sample group.

Deletion of outliers for one variable can significantly affect the frequency distribution of the other variables. Thus, straight lines or continuous distributions can be difficult to achieve. Through iteration, most of the outliers can be deleted. It is important that outliers are deleted from the reference group since the inclusion of any outliers distorts the means and variance-covariances of the sample population and subsequently affects any statistically based classification scheme. It would be almost impossible to delete every atypical sample, however, this approach provides an initial

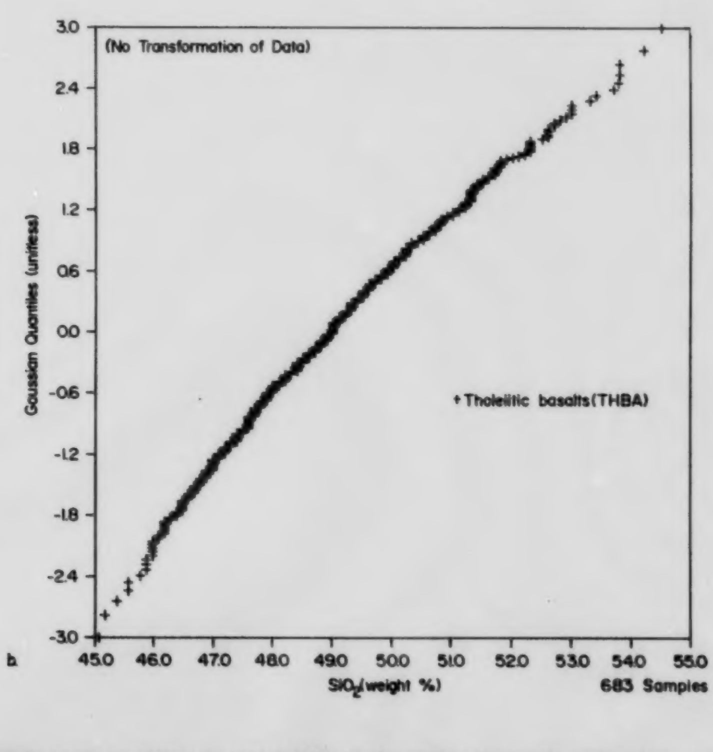

**Figure 28.6.** a) Quantile-Quantile (Q-Q) plot of $SiO_2$ for 1227 tholeiitic basalts (THBA) showing the outliers. Notice how the presence of outliers distorts the shape of the trend of the data. b) Q-Q plot of $SiO_2$ for 683 tholeiitic basalts (THBA) with the outliers removed. The trend of the data follows a straight line without the outliers.

set of samples by which typicality of samples can be measured.

In the case of irregular, or poorly defined trends in the data, the frequency distribution must be accepted as a limiting fact on the ability to describe the data. Experience in this study has shown that such distributions occur when the reference group is composed of more than one population. This reflects back on the initial classification used to set up the reference groups and suggests that the classification has not subdivided all of the samples into discrete populations. This problem was noted in the Q-Q plots that describe the frequency distribution of the fractionated tholeiitic rocks. It was found that the tholeiitic dacites and rhyolites were bimodal. This will be discussed *in* Application of Allocation Procedures to PETROCH Samples.

In addition to the use of Q-Q plots, the use of chi-square ($\chi^2$) plots as described by Garrett (1989b) assists in detecting multivariate outliers. This approach considers the distance of each sample from the compositional centroid of the reference group in multivariate space. This distance, the Mahalanobis distance, is then plotted against the theoretical values of the $\chi^2$ distribution. Outliers are clearly identified by their large distances compared to the expected $\chi^2$ distances. Prior to determining the Mahalanobis distance, the individual variables that define the data should be transformed to normality. This can be achieved by fitting power transformations to the variables where necessary. Samples with overly large Mahalanobis distances are considered to be outliers. The rejection criteria is based on robust estimation theory (cf., Gnanadesikan 1977; Hampel et al. 1986).

Figure 28.7a shows a $\chi^2$ plot for 715 samples of the iron-rich tholeiite (THFE) reference group prior to the elimination of the outliers using Q-Q plots. The samples plot as a curved line due to the presence of atypical samples. Figure 28.7b shows 258 samples of the same reference groups after the elimination of outliers using Q-Q plots. In this case, the samples form a straight line, which corresponds to a multivariate normal population.

Column B of Table 28.2 shows the results of trimming the reference groups after examining the Q-Q and $\chi^2$ plots. The numbers of samples within each group have been reduced substantially. Most samples were eliminated based on atypical abundances of $K_2O$, $Na_2O$ and LOI. Of the initial komatiitic rocks, only 25% remain in the komatiitic clan for a total of 155 samples. For the tholeiitic rocks, roughly 45% of the samples have been retained (1112 samples). For the calc-alkalic rocks, 43% of the samples have been retained (578 samples) within the reference groups.

## CHEMICAL VARIATION AND CHARACTERIZATION OF THE REFERENCE GROUPS

Table 28.3 provides statistical descriptions of the trimmed reference groups in tabular format. Each table lists the

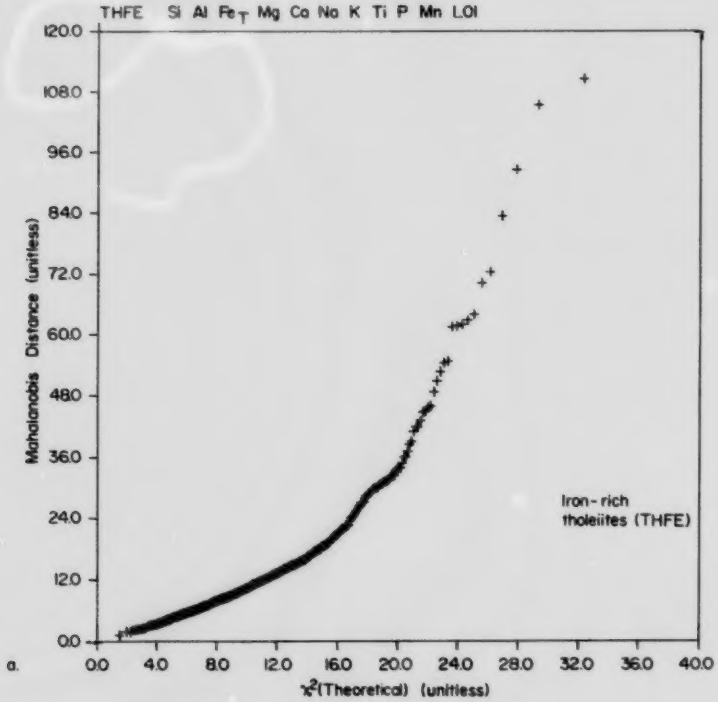

**Figure 28.7. a)** $\chi^2$ plot of 715 iron-rich tholeiites (THFE) prior to removing outliers. The samples form a curved line, which indicates that the sample population is not multivariate normal. **b)** $\chi^2$ plot of 258 iron-rich tholeiites (THFE) after the removal of the outliers The samples plot along a straight line, which indicates that the population is close to multivariate normal.

minimum, maximum, median, mean and standard deviations for each major element oxide. These values provide a brief tabular description of the data and a useful range of values by which unknown samples can be compared. These statistics are based on anhydrous values for $SiO_2$, $Al_2O_3$, $Fe_2O_3$, FeO, MgO, CaO, $Na_2O$, $K_2O$, $TiO_2$, $P_2O_5$ and MnO. The statistics for $CO_2$, S, $H_2O^+$ and LOI are determined from the original entries in the PETROCH data base. These tables are intended to provide an initial estimate of least altered chemical compositions for the major volcanic rock types of the Superior Province. Future work may result in modifications of these values. These tables can be used as a basis for comparison with other rock type definitions (e.g., Irvine and Baragar 1971; Basaltic Volcanism Study Project 1981).

## Reference Groups Plotted onto TAS, AFM and AFTiM Variation Diagrams

Figures 28.8, 28.9 and 28.10 show only the samples (1845) that were retained for the reference groups. Figure 28.8 shows the reference groups plotted on the $Na_2O+K_2O$ versus $SiO_2$ (TAS) diagram. Figures 28.9 and 28.10 show the 3 reference suites plotted on the $Na_2O+K_2O$-$FeO^T$-MgO (AFM) diagram and the $AlO_{1.5}$-$FeO_{1.5}$+FeO+$TiO_2$-MgO (AFTiM) molecular proportion diagram of Jensen (1976).

Figure 28.8 shows the komatiitic and mafic tholeiitic rocks form distinct clouds of groups. The calc-alkalic rocks display a dispersed trend following the increase of $SiO_2$ and $Na_2O+K_2O$. Relative to Figure 28.3, the dispersion of the samples is notably less; the samples fall within the limits of the compositional ranges defined by the TAS classification (LeMaitre 1989).

Figure 28.9 shows the komatiitic rocks form a distinct group of samples along the $FeO^T$-MgO side of the ternary diagram. The mafic tholeiitic rocks display 3 relatively distinct clouds of points that define the magnesium-rich, iron-rich and normal tholeiitic basalts. The calc-alkalic rocks display a broad dispersion of points extending along the tholeiitic–calc-alkalic dividing line.

Figure 28.10 shows the mafic tholeiitic rocks display 3 distinct clouds of samples. The fractionated tholeiitic rocks show a dispersed range of compositions along the tholeiitic–calc-alkalic dividing line. The calc-alkalic samples also show a broad trend of compositional variation along the tholeiitic–calc-alkalic dividing line. As in the case of the AFM diagram, the classification of fractionated tholeiitic and calc-alkalic rocks should be based on the environment of association with other rocks in the sequence.

## Principal Components Analysis Applied to the Geochemistry of the Reference Groups

The characterization of the chemistry of the reference groups can be carried out using conventional variation

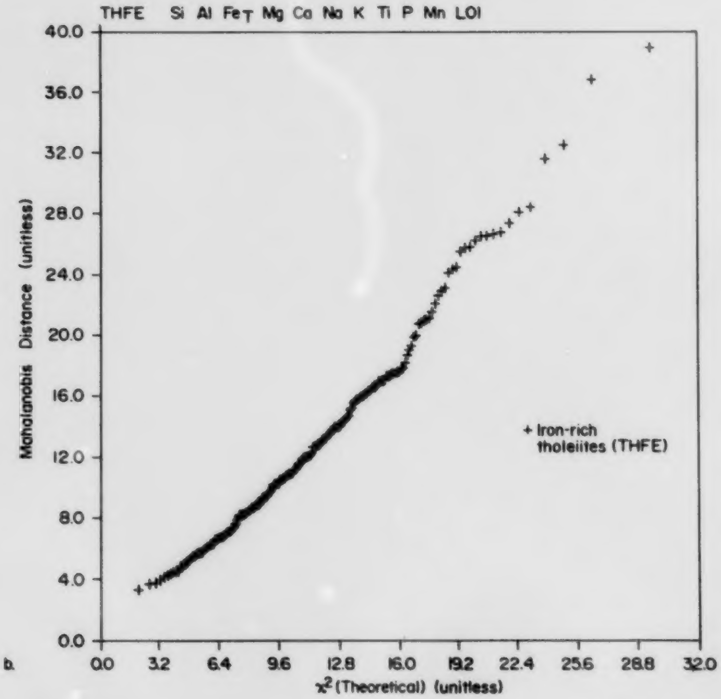

**Figure 28.7.** a) $\chi^2$ plot of 715 iron-rich tholeiites (THFE) prior to removing outliers. The samples form a curved line, which indicates that the sample population is not multivariate normal. b) $\chi^2$ plot of 258 iron-rich tholeiites (THFE) after the removal of the outliers The samples plot along a straight line, which indicates that the population is close to multivariate normal.

**Table 28.3a.** Summary statistics for least altered reference group: PETROCH data base group 1, ultramafic komatiite (KOUM).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 83 | 43.71 | 51.38 | 47.54 | 47.25 | 1.54 |
| $Al_2O_3$ | 83 | 3.60 | 10.19 | 6.22 | 6.13 | 1.77 |
| $FeO^T$ | 83 | 8.64 | 12.96 | 10.91 | 11.01 | .96 |
| MgO | 83 | 18.04 | 33.28 | 26.85 | 26.52 | 3.19 |
| CaO | 83 | 4.54 | 12.89 | 7.35 | 7.57 | 1.67 |
| $Na_2O$ | 83 | .01 | .47 | .19 | .22 | .13 |
| $K_2O$ | 83 | .01 | .26 | .03 | .04 | .04 |
| $TiO_2$ | 83 | .19 | .58 | .38 | .38 | .08 |
| $P_2O_5$ | 83 | .01 | .12 | .02 | .03 | .02 |
| MnO | 83 | .11 | .34 | .19 | .20 | .04 |
| $CO_2$ | 41 | .05 | 2.34 | .26 | .49 | .56 |
| S | 43 | .01 | 1.01 | .02 | .06 | .15 |
| $H_2O^+$ | 37 | 3.22 | 8.96 | 5.41 | 5.62 | 1.39 |
| LOI | 83 | 3.10 | 7.89 | 6.15 | 6.20 | 1.17 |

**Table 28.3b.** Summary statistics for least altered reference group: PETROCH data base group 2, basaltic komatiite (KOBA).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 72 | 47.58 | 53.82 | 49.52 | 49.81 | 1.53 |
| $Al_2O_3$ | 72 | 6.25 | 14.61 | 11.60 | 11.12 | 1.86 |
| $FeO^T$ | 72 | 9.72 | 12.60 | 11.12 | 11.03 | .71 |
| MgO | 72 | 10.20 | 20.08 | 14.18 | 14.92 | 2.97 |
| CaO | 72 | 7.61 | 14.43 | 10.26 | 10.26 | 1.81 |
| $Na_2O$ | 72 | .53 | 2.26 | 1.17 | 1.28 | .46 |
| $K_2O$ | 72 | .01 | .50 | .13 | .15 | .09 |
| $TiO_2$ | 72 | .44 | .82 | .58 | .59 | .09 |
| $P_2O_5$ | 72 | .01 | .13 | .03 | .04 | .03 |
| MnO | 72 | .15 | .26 | .20 | .20 | .02 |
| $CO_2$ | 58 | .05 | .91 | .15 | .25 | .20 |
| S | 61 | .01 | .23 | .02 | .04 | .05 |
| $H_2O^+$ | 39 | .70 | 4.50 | 2.22 | 2.48 | 1.07 |
| LOI | 72 | .60 | 5.50 | 2.80 | 2.77 | 1.34 |

**Table 28.3c.** Summary statistics for least altered reference group: PETROCH data base group 3, iron-rich tholeiite (THFE).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 258 | 48.25 | 55.38 | 51.00 | 51.19 | 1.64 |
| $Al_2O_3$ | 258 | 11.98 | 15.78 | 13.85 | 13.93 | .76 |
| $FeO^T$ | 258 | 11.88 | 17.13 | 14.67 | 14.65 | 1.17 |
| MgO | 258 | 2.47 | 7.00 | 5.58 | 5.46 | .84 |
| CaO | 258 | 5.31 | 14.48 | 9.12 | 9.12 | 1.53 |
| $Na_2O$ | 258 | 1.47 | 3.88 | 2.40 | 2.50 | .57 |
| $K_2O$ | 258 | .05 | 1.02 | .23 | .28 | .20 |
| $TiO_2$ | 258 | 1.03 | 2.41 | 1.66 | 1.68 | .30 |
| $P_2O_5$ | 258 | .01 | .27 | .12 | .13 | .06 |
| MnO | 258 | .17 | .32 | .24 | .24 | .03 |
| $CO_2$ | 226 | .02 | 3.68 | .26 | .49 | .57 |
| S | 233 | .01 | 1.34 | .09 | .12 | .15 |
| $H_2O^+$ | 170 | .31 | 4.56 | 2.27 | 2.25 | .93 |
| LOI | 258 | .50 | 3.90 | 2.00 | 1.94 | .94 |

**Table 28.3d.** Summary statistics for least altered reference group: PETROCH data base group 4, magnesium-rich tholeiite (THMG).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 74 | 47.39 | 54.53 | 50.72 | 51.14 | 1.64 |
| $Al_2O_3$ | 74 | 12.70 | 17.38 | 15.12 | 14.98 | 1.08 |
| $FeO^T$ | 74 | 8.58 | 11.17 | 10.08 | 10.00 | .66 |
| MgO | 74 | 8.12 | 11.16 | 9.19 | 9.28 | .70 |
| CaO | 74 | 7.40 | 14.24 | 10.84 | 10.73 | 1.41 |
| $Na_2O$ | 74 | 1.28 | 3.12 | 2.09 | 2.12 | .47 |
| $K_2O$ | 74 | .02 | .72 | .17 | .24 | .18 |
| $TiO_2$ | 74 | .50 | 1.10 | .73 | .74 | .13 |
| $P_2O_5$ | 74 | .01 | .20 | .05 | .05 | .03 |
| MnO | 74 | .14 | .22 | .18 | .18 | .02 |
| $CO_2$ | 67 | .07 | 3.05 | .29 | .63 | .71 |
| S | 67 | .01 | .25 | .03 | .05 | .05 |
| $H_2O^+$ | 45 | .78 | 4.30 | 2.22 | 2.22 | .90 |
| LOI | 74 | .60 | 6.00 | 2.30 | 2.48 | 1.26 |

**Table 28.3e.** Summary statistics for least altered reference group: PETROCH data base group 5, tholeiitic basalt (THBA).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 683 | 47.11 | 56.29 | 50.60 | 50.84 | 1.57 |
| $Al_2O_3$ | 683 | 13.20 | 16.73 | 15.07 | 15.08 | .73 |
| $FeO^T$ | 683 | 9.63 | 14.50 | 11.88 | 11.91 | 1.08 |
| MgO | 683 | 5.46 | 8.98 | 7.29 | 7.26 | .79 |
| CaO | 683 | 5.75 | 13.26 | 10.49 | 10.34 | 1.31 |
| $Na_2O$ | 683 | 1.34 | 3.77 | 2.33 | 2.39 | .56 |
| $K_2O$ | 683 | .02 | .74 | .17 | .21 | .15 |
| $TiO_2$ | 683 | .55 | 1.90 | 1.00 | 1.05 | .24 |
| $P_2O_5$ | 683 | .01 | .23 | .07 | .08 | .04 |
| MnO | 683 | .14 | .37 | .21 | .21 | .03 |
| $CO_2$ | 619 | .02 | 3.12 | .29 | .49 | .51 |
| S | 626 | .01 | .77 | .05 | .07 | .08 |
| $H_2O^+$ | 408 | .31 | 4.18 | 2.11 | 2.11 | .96 |
| LOI | 683 | .50 | 5.40 | 2.10 | 2.11 | 1.14 |

**Table 28.3f.** Summary statistics for least altered reference group: PETROCH data base group 6, calc-alkalic basalt (CABA).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 265 | 49.75 | 64.71 | 57.84 | 57.56 | 2.87 |
| $Al_2O_3$ | 265 | 13.47 | 19.48 | 16.16 | 16.21 | 1.12 |
| $FeO^T$ | 265 | 5.09 | 12.15 | 7.43 | 7.77 | 1.23 |
| MgO | 265 | 3.19 | 7.10 | 5.28 | 5.24 | .86 |
| CaO | 265 | 3.61 | 11.79 | 6.78 | 7.07 | 1.57 |
| $Na_2O$ | 265 | 2.39 | 5.44 | 3.56 | 3.72 | .68 |
| $K_2O$ | 265 | .31 | 1.55 | .65 | .72 | .31 |
| $TiO_2$ | 265 | .45 | 1.58 | .91 | .94 | .23 |
| $P_2O_5$ | 265 | .01 | .32 | .14 | .14 | .07 |
| MnO | 265 | .07 | .26 | .14 | .14 | .04 |
| $CO_2$ | 242 | .03 | 6.04 | .47 | 1.11 | 1.37 |
| S | 238 | .01 | 2.12 | .01 | .06 | .19 |
| $H_2O^+$ | 169 | .55 | 3.84 | 2.48 | 2.38 | .82 |
| LOI | 265 | .70 | 7.60 | 2.80 | 3.20 | 1.72 |

**Table 28.3g.** Summary statistics for least altered reference group: PETROCH data base group 7, calc-alkalic andesite (CAAN).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 138 | 56.37 | 67.59 | 61.59 | 61.45 | 2.52 |
| $Al_2O_3$ | 138 | 14.57 | 18.88 | 16.54 | 16.58 | .92 |
| $FeO^T$ | 138 | 4.20 | 8.22 | 5.88 | 5.95 | .89 |
| MgO | 138 | 1.95 | 5.26 | 3.62 | 3.54 | .74 |
| CaO | 138 | 3.88 | 9.48 | 6.00 | 6.10 | 1.11 |
| $Na_2O$ | 138 | 3.04 | 5.05 | 4.05 | 4.05 | .53 |
| $K_2O$ | 138 | .26 | 2.05 | .86 | .96 | .48 |
| $TiO_2$ | 138 | .35 | 1.27 | .75 | .77 | .21 |
| $P_2O_5$ | 138 | .03 | .34 | .13 | .14 | .06 |
| MnO | 138 | .06 | .16 | .10 | .11 | .02 |
| $CO_2$ | 126 | .06 | 4.29 | .32 | .84 | .96 |
| S | 124 | .01 | 1.52 | .02 | .06 | .15 |
| $H_2O^+$ | 89 | .40 | 3.85 | 2.07 | 1.94 | .73 |
| LOI | 138 | .20 | 6.00 | 2.20 | 2.42 | 1.35 |

**Table 28.3h.** Summary statistics for least altered reference group: PETROCH data base group 8, calc-alkalic dacite (CADA).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 113 | 61.70 | 73.86 | 66.95 | 67.17 | 2.75 |
| $Al_2O_3$ | 113 | 12.82 | 18.64 | 15.93 | 16.01 | 1.22 |
| $FeO^T$ | 113 | 2.41 | 5.39 | 3.67 | 3.76 | .70 |
| MgO | 113 | .72 | 2.99 | 1.83 | 1.84 | .57 |
| CaO | 113 | 2.14 | 6.77 | 4.18 | 4.22 | .88 |
| $Na_2O$ | 113 | 3.19 | 6.27 | 4.66 | 4.67 | .76 |
| $K_2O$ | 113 | .27 | 3.31 | 1.30 | 1.40 | .68 |
| $TiO_2$ | 113 | .23 | .87 | .48 | .51 | .14 |
| $P_2O_5$ | 113 | .01 | .29 | .11 | .12 | .06 |
| MnO | 113 | .02 | .14 | .06 | .07 | .02 |
| $CO_2$ | 105 | .06 | 3.48 | .61 | 1.14 | 1.03 |
| S | 103 | .01 | .58 | .01 | .06 | .11 |
| $H_2O^+$ | 70 | .50 | 2.71 | 1.35 | 1.37 | .49 |
| LOI | 113 | .30 | 4.40 | 1.90 | 2.19 | 1.16 |

**Table 28.3i.** Summary statistics for least altered reference group: PETROCH data base group 9, calc-alkalic rhyolite (CARH).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 62 | 68.13 | 76.24 | 72.25 | 72.18 | 2.15 |
| $Al_2O_3$ | 62 | 11.97 | 17.59 | 15.09 | 15.02 | 1.24 |
| $FeO^T$ | 62 | .60 | 3.41 | 1.93 | 1.96 | .60 |
| MgO | 62 | .26 | 1.55 | .67 | .74 | .32 |
| CaO | 62 | 1.19 | 4.84 | 2.85 | 2.93 | .82 |
| $Na_2O$ | 62 | 3.69 | 6.30 | 4.88 | 4.94 | .70 |
| $K_2O$ | 62 | .51 | 3.34 | 1.63 | 1.69 | .66 |
| $TiO_2$ | 62 | .16 | .71 | .34 | .35 | .13 |
| $P_2O_5$ | 62 | .01 | .16 | .06 | .07 | .04 |
| MnO | 62 | .01 | .09 | .04 | .05 | .02 |
| $CO_2$ | 53 | .08 | 3.74 | 1.18 | 1.41 | .89 |
| S | 48 | .01 | .82 | .01 | .07 | .16 |
| $H_2O^+$ | 37 | .11 | 1.37 | .78 | .78 | .29 |
| LOI | 62 | .20 | 4.30 | 1.80 | 1.97 | .94 |

**Table 28.3j.** Summary statistics for least altered reference group: PETROCH data base group 10, tholeiitic andesite (THAN).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 49 | 48.83 | 57.95 | 52.96 | 52.94 | 2.14 |
| $Al_2O_3$ | 49 | 14.44 | 19.50 | 16.31 | 16.32 | 1.02 |
| $FeO^T$ | 49 | 8.31 | 13.19 | 11.19 | 11.24 | 1.04 |
| MgO | 49 | 2.31 | 5.29 | 3.73 | 3.76 | .62 |
| CaO | 49 | 4.95 | 14.13 | 9.16 | 9.44 | 2.29 |
| $Na_2O$ | 49 | 2.08 | 4.88 | 2.98 | 3.22 | .80 |
| $K_2O$ | 49 | .03 | .78 | .28 | .33 | .19 |
| $TiO_2$ | 49 | 1.06 | 2.77 | 1.47 | 1.63 | .45 |
| $P_2O_5$ | 49 | .01 | .44 | .11 | .15 | .09 |
| MnO | 49 | .14 | .37 | .25 | .25 | .06 |
| $CO_2$ | 43 | .12 | 5.72 | 1.51 | 1.92 | 1.38 |
| S | 45 | .01 | .63 | .06 | .09 | .10 |
| $H_2O^+$ | 26 | .16 | 3.74 | 2.45 | 2.11 | 1.00 |
| LOI | 49 | .50 | 6.90 | 3.10 | 3.32 | 1.86 |

**Table 28.3k.** Summary statistics for least altered reference group: PETROCH data base group 11, tholeiitic dacite (THDA).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 25 | 58.28 | 71.53 | 63.64 | 64.15 | 3.23 |
| $Al_2O_3$ | 25 | 12.20 | 18.10 | 14.63 | 15.10 | 1.46 |
| $FeO^T$ | 25 | 6.00 | 8.49 | 7.20 | 7.25 | .76 |
| MgO | 25 | .85 | 2.88 | 1.77 | 1.83 | .55 |
| CaO | 25 | 2.59 | 8.57 | 4.03 | 4.46 | 1.30 |
| $Na_2O$ | 25 | 2.74 | 5.68 | 4.34 | 4.31 | .73 |
| $K_2O$ | 25 | .76 | 2.27 | 1.16 | 1.21 | .36 |
| $TiO_2$ | 25 | .50 | 1.18 | .91 | .90 | .18 |
| $P_2O_5$ | 25 | .09 | .45 | .19 | .23 | .10 |
| MnO | 25 | .07 | .26 | .14 | .15 | .05 |
| $CO_2$ | 23 | .14 | 3.82 | .65 | 1.04 | .96 |
| S | 23 | .01 | 1.34 | .02 | .10 | .28 |
| $H_2O^+$ | 16 | .41 | 2.48 | 1.58 | 1.45 | .61 |
| LOI | 25 | .57 | 5.20 | 1.60 | 2.21 | 1.35 |

**Table 28.3l.** Summary statistics for least altered reference group: PETROCH data base group 12, tholeiitic rhyolite (THRH).

| Element Oxide | No. of Samples | Min. | Max. | Median | Mean | Std. Dev. |
|---|---|---|---|---|---|---|
| $SiO_2$ | 23 | 64.89 | 77.89 | 68.15 | 69.71 | 3.66 |
| $Al_2O_3$ | 23 | 10.89 | 16.74 | 14.65 | 14.55 | 1.61 |
| $FeO^T$ | 23 | 3.12 | 6.45 | 4.35 | 4.55 | .96 |
| MgO | 23 | .32 | 1.56 | .74 | .89 | .35 |
| CaO | 23 | 1.34 | 4.88 | 3.21 | 3.17 | .96 |
| $Na_2O$ | 23 | 2.73 | 6.33 | 4.31 | 4.46 | .87 |
| $K_2O$ | 23 | .90 | 3.16 | 1.42 | 1.59 | .66 |
| $TiO_2$ | 23 | .27 | 1.40 | .49 | .60 | .27 |
| $P_2O_5$ | 23 | .03 | .30 | .10 | .12 | .08 |
| MnO | 23 | .05 | .14 | .08 | .09 | .03 |
| $CO_2$ | 22 | .06 | 3.72 | .59 | 1.10 | 1.12 |
| S | 21 | .01 | .59 | .02 | .07 | .13 |
| $H_2O^+$ | 20 | .29 | 2.09 | .99 | 1.01 | .44 |
| LOI | 23 | .61 | 4.10 | 1.40 | 1.86 | 1.03 |

diagrams as mentioned above, but such diagrams display only 2, 3, and sometimes 4 elements. This fails to give insight into the behaviour of all of the elements simultaneously and thus can be misleading. A desirable goal is to come up with a numerical and visual procedure that provides insight into all of the elements simultaneously. The most commonly used methods to do this are components analysis methods. Methods such as principal components analysis (PCA) reduce the number of elements that define the primary geochemical variation. Thus, the data can be projected onto a reduced number of "axes" that represent linear combinations of the elements, based on the maximum variances of the variables, such that only a few "components" are required to explain most of the geochemical variation. These techniques are explained in Davis (1986) and Jöreskog et al. (1976). These methods also provide a visual means of examining the element associations by projecting combinations of the component axes onto two-dimensional planes. The component scores of the elements (relative relationships of the elements) and the scores of the samples (the relative relationships of the samples) can be plotted and viewed to assess the geochemical associations and affinities. Conventional PCA procedures (cf., Jöreskog et al. 1976; Davis 1986) have the drawback that they must be used in either Q-mode (sample based) or R-mode (variable based). Thus, diagrams of the factor scores projected onto a two-dimensional plane can only show the relationships of the samples relative to each other or the relationships of the elements relative to each other, but not the relationships between the samples and the elements on the same diagram. More recently, a simultaneous RQ-mode solution to PCA has been presented (Zhou et al. 1983; Davis 1986, p.594) and has been used successfully to describe the variation of the chemistry of the reference suites.

Aitchison (1986, p.186) discussed the problems associated with the use of so-called crude principal components analysis applied to constant sum (closed) data. The difficulty in the interpretation of the crude covariance or correlation matrix of compositional data creates difficulty in assessing the multivariate relationships of the samples and variables. Aitchison suggested that the use of multiplicative logratio transformations is the correct way to investigate the multivariate nature of compositional data.

RQ-mode logcontrast principal components analysis was applied to the 3 reference groups for the following major element oxides: $SiO_2$, $Al_2O_3$, $FeO^T$, MgO, CaO, $Na_2O$, $K_2O$, $TiO_2$, $P_2O_5$, MnO and LOI. The multiplicative logratio transform was used as outlined in the section entitled Compositional Data, The Closure Problem and the Use of Logratio Transformations. However, it was found that many $P_2O_5$ analyses were equal to zero and a replacement value of 0.005 had been substituted according to the procedure outlined in Appendix 1. The resulting logratio transform resulted in exceptionally large negative values for these samples and thus created a large variation in the $P_2O_5$ distribution. This distortion in scale produced a bimodal

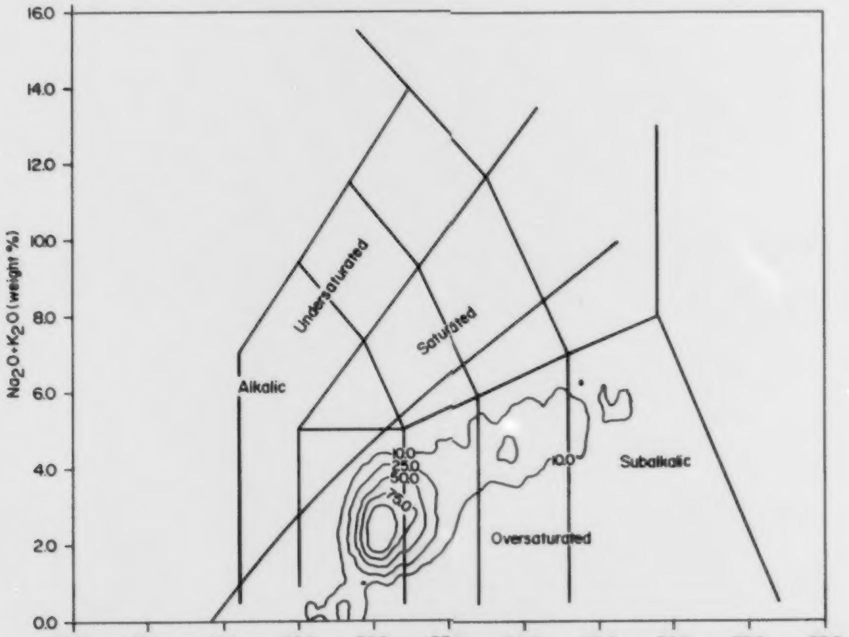

**Figure 28.8.** Total Alkalis-Silica (TAS) diagram (*modified from* LeMaitre 1989) for 1845 "least altered" samples obtained from procedures outlined in the text. Note the lesser degree of dispersion in comparison with Figure 28.3. The contour lines represent densities per unit cell in the figure. The figure was divided into 100 cells along the x-axis, 80 cells along the y-axis. Contour intervals are: 10, 25, 50, 75, 100, 150, 200 and 250 points per square unit.

multivariate population that was difficult to interpret. Such scale distortions are a problem and at this time not fully resolved. It was reasoned that the $P_2O_5$ analyses should not have this degree of significance and formed the basis for rejecting samples that contained zero abundances for $P_2O_5$ as discussed in the section entitled Data Selection.

The results of the analysis are shown in Table 28.4. The table lists the eigenvalues (λ), variation accounted for by each eigenvalue (trace), the cumulative trace, the R-mode loadings of the variables, and the relative contributions of the variables. The R-mode loadings provide information regarding the relative relationships of the variables within each component. These relationships are best displayed graphically (Figure 28.11). The relative contributions provide information with respect to how much each variable contributes to each component and thus gives a measure of how significant a variable is within each component. The primary aim of the analysis is to describe the most significant chemical variation in as few variables as possible. In this case, the first 3 principal components account for more than 81% of the data variation. Table 28.4 also shows the R-mode loadings of the variables in which the first component (50% of the data variation) shows an inverse relationship of $SiO_2$, $Al_2O_3$, $Na_2O$ and $K_2O$ with MgO, $FeO^T$, CaO, $TiO_2$, MnO and LOI. The relative contributions show that most of the $FeO^T$, MnO, MgO, $K_2O$, $Na_2O$ and CaO variation are accounted for in the first component. The relative contributions of the second component (17% of the data variation) show that $TiO_2$, $SiO_2$ and LOI are the dominant variables. The third component (14% of the data variation) primarily accounts for $Al_2O_3$, LOI, $P_2O_5$ and CaO and may reflect a combination of plagioclase, alteration and excessive $P_2O_5$ values. The remaining components account for less than 20% of the variation in the data. Some of the lesser components may represent specific combinations of elements representing specific rock types while others may be due to random effects.

Figure 28.11 shows the multivariate relationships graphically. The sample and variable loadings are plotted onto the first 2 component axes. The relationships of the samples show that the komatiitic suite forms a distinct cloud along the negative end of the PC1 axis. The corresponding relative position of the variables shows that these samples are magnesium enriched. Similarly, the tholeiitic rocks plot close to the origin associated with iron, manganese, calcium and titanium enrichment. The calc-alkalic rocks plot along the positive side of the PC1 axis and show an association with sodium and potassium enrichment. The dispersion of

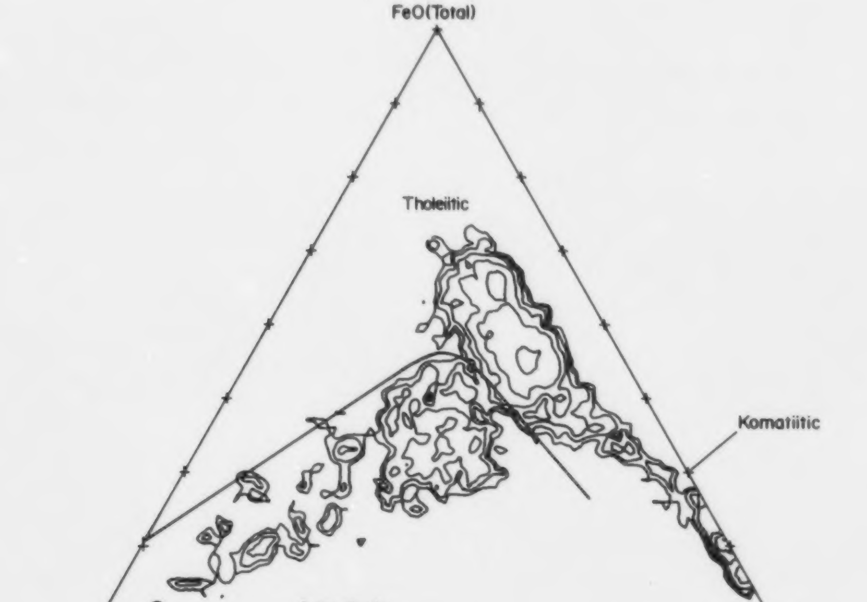

**Figure 28.9.** AFM diagram (*modified from* Irvine and Baragar 1971) for 1845 "least altered" samples obtained from procedures outlined in the text. The komatiitic, tholeiitic and calc-alkalic trends are clearly observed in this plot. Contours are defined as units of standard deviation from the expected sum of a random distribution using a 1% search triangle. Contour maximum represents 40.636 points at $Na_2O+K_2O$ = 12 MgO = 33 $FeO^T$ = 55.

samples along the PC2 axis indicates the relative dispersion due to LOI enrichment as indicated in Table 28.4. Thus, PCA is able to show most of the compositional variation of the reference groups in a single plot. The relative relationships of the sample and the variables are expressed in this diagram and more than 67% of the variation of the data is accounted for. This is far more informative than any of the ternary or binary discriminating diagrams.

If altered samples were to be included in the reference groups, they would plot as isolated samples away from the main group of samples. As well, the sample would plot in proximity to the major element oxide in which it would be relatively enriched. Examples of this are shown in Grunsky (1986).

## KOMATIITIC REFERENCE GROUP

Logcontrast principal components analysis was applied to the komatiitic reference suite and the results are shown in Table 28.5. The first 5 components account for 90% of the total variation of the data. The first and second components account for 50% and 16% of the data variation, respectively. Examination of the R-mode loadings for the first component indicates that the most significant variables are MgO, $FeO^T$, $SiO_2$, LOI and $Na_2O$. $SiO_2$, MgO, CaO, $P_2O_5$, MnO, LOI and $FeO^T$ are inversely associated with $Al_2O_3$, $Na_2O$, $K_2O$ and $TiO_2$. The second component shows CaO as the dominant variable and represents sodium-enriched komatiitic rocks which are probably the result of metasomatism. Figure 28.12 shows a plot of the sample and variable scores plotted onto the first 2 component axes. Ultramafic komatiitic rocks (KOUM) are associated with the positive side of the PC1 axis and show relative enrichment in MgO, LOI and $SiO_2$. The association of $SiO_2$ with MgO and LOI reflects $SiO_2$ abundance relative to basaltic komatiites that are $TiO_2$ enriched with correspondingly less amounts of $SiO_2$. The basaltic komatiites (KOBA) along the negative side of the PC1 axis show a relative enrichment in $Al_2O_3$, $TiO_2$, $K_2O$ and $Na_2O$. Each rock type (KOUM, KOBA) is contoured so that the relative densities and overlap can be observed.

The third component (11% of the data variation) reflects enrichment in $P_2O_5$, $TiO_2$ and $K_2O$. This component represents samples that have individual enrichment and that reflect some of the non-normal characteristics of the data.

Three components account for 80% of the variation of the data and characterize the most important relationships between the samples and major element oxides.

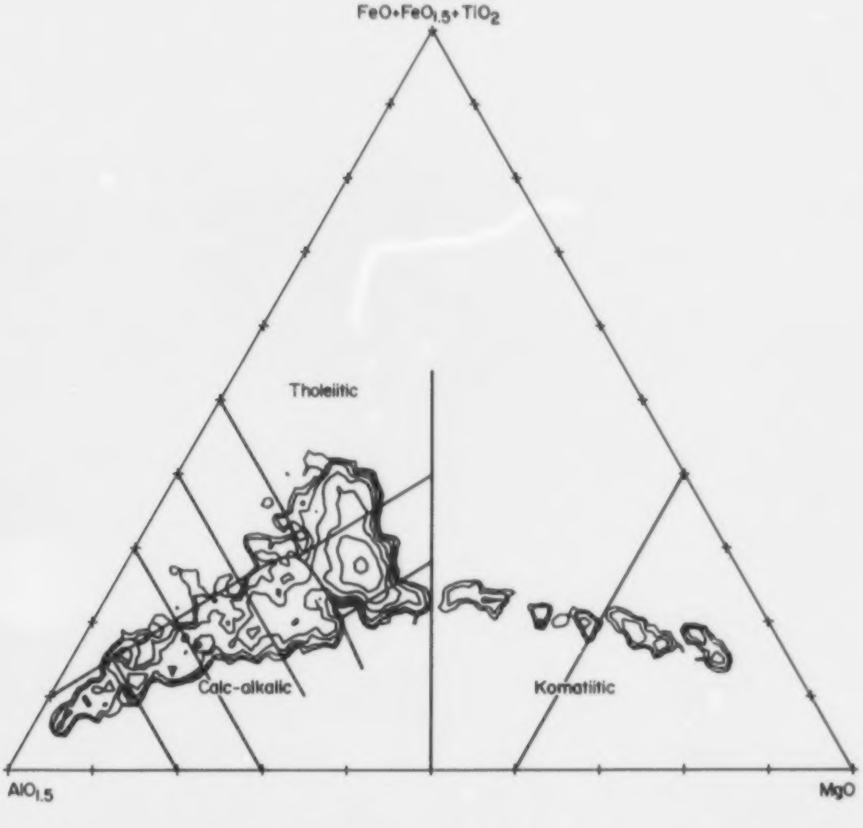

**Figure 28.10.** AFTiM diagram (*after* Jensen 1976) for 1845 "least altered" samples obtained from procedures outlined in the text. The komatiitic, tholeiitic and calc-alkalic trends are clearly observed in this plot. Contours are defined as units of standard deviation from the expected sum of a random distribution using a 1% search triangle. Contour maximum represents 71.691 points at Al=44, Fe+Ti=28, Mg=28.

**Table 28.4.** Multiplicative logratio principal components analysis: All PETROCH reference groups. [KOM/THOL/CALC] Samples: 1845

| Component | Eigenvalue | % Trace | Σ% Trace |
|---|---|---|---|
| 1 | 5.5107 | 50.0977 | 50.0977 |
| 2 | 1.9098 | 17.3617 | 67.4594 |
| 3 | 1.5069 | 13.6987 | 81.1581 |
| 4 | 0.5963 | 5.4206 | 86.5787 |
| 5 | 0.5586 | 5.0784 | 91.6571 |
| 6 | 0.3644 | 3.3126 | 94.9697 |
| 7 | 0.1907 | 1.7333 | 96.7030 |
| 8 | 0.1815 | 1.6497 | 98.3528 |
| 9 | 0.1230 | 1.1184 | 99.4712 |
| 10 | 0.0582 | 0.5288 | 100.0000 |

| Component* | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| **Principal Component Loadings** | | | | | |
| Si | 0.4384 | -0.6634 | 0.3418 | -0.0735 | 0.3025 |
| Al | 0.5721 | -0.1122 | 0.7161 | 0.1674 | 0.1930 |
| $Fe^T$ | -0.9518 | 0.1459 | 0.1110 | -0.0104 | 0.0234 |
| Mg | -0.8684 | -0.3254 | -0.0436 | -0.2048 | 0.0015 |
| Ca | -0.8107 | -0.0882 | 0.4487 | -0.1090 | 0.0678 |
| Na | 0.8114 | 0.2937 | 0.2952 | 0.2099 | -0.0856 |
| K | 0.8667 | 0.0307 | -0.0326 | -0.2312 | -0.3929 |
| Ti | -0.4367 | 0.7546 | 0.1066 | 0.3537 | 0.0652 |
| P | 0.4738 | 0.4721 | -0.4848 | -0.2256 | 0.5060 |
| Mn | -0.9138 | 0.1070 | 0.1415 | -0.0105 | -0.0417 |
| LOI | -0.1194 | -0.6567 | -0.5541 | 0.4851 | 0.0321 |
| **Relative Contributions (%)** | | | | | |
| Si | 19.2235 | 44.0067 | 11.6841 | 0.5397 | 9.1494 |
| Al | 32.7326 | 1.2584 | 51.2756 | 2.8038 | 3.7233 |
| $Fe^T$ | 90.5852 | 2.1281 | 1.2322 | 0.0108 | 0.0550 |
| Mg | 75.4135 | 10.5881 | 0.1905 | 4.1957 | 0.0002 |
| Ca | 65.7188 | 0.7773 | 20.1330 | 1.1875 | 0.4599 |
| Na | 65.8409 | 8.6260 | 8.7170 | 4.4063 | 0.7323 |
| K | 75.1088 | 0.0945 | 0.1061 | 5.3444 | 15.4386 |
| Ti | 19.0715 | 56.9396 | 1.1362 | 12.5124 | 0.4255 |
| P | 22.4506 | 22.2905 | 23.5067 | 5.0878 | 25.6013 |
| Mn | 83.5025 | 1.1441 | 2.0022 | 0.0111 | 0.1737 |
| LOI | 1.4267 | 43.1255 | 30.7016 | 23.5276 | 0.1032 |

Major element oxides are denoted by their elements only.

* *Only the first 5 components are listed.*

## THOLEIITIC REFERENCE GROUP

Table 28.6 shows the results of logcontrast principal components analysis applied to the tholeiitic reference group. The first 5 components account for more than 87% of the total variation in the data. The relative contributions of the variables indicate that the first component (36% of the data variation) accounts for most of the variation of CaO, MgO, $FeO^T$, MnO and $K_2O$. The second component (23% of the data variation) accounts for most of the variation of $SiO_2$, $Al_2O_3$ and $TiO_2$. The third component (12% of the data variation) accounts for most of the variation of LOI.

For the first component, the R-mode loadings of the variable show that CaO, MgO, $FeO^T$ and MnO are inversely associated with $K_2O$ and $Na_2O$, which reflects the primary compositional difference between the mafic and felsic tholeiitic rocks. The loadings for the second component show that $SiO_2$ and $Al_2O_3$ are inversely associated with $TiO_2$. This component accounts for the main difference between the high iron tholeiites from the magnesium tholeiites. The third component indicates that LOI is inversely associated with all of the major element oxides with the exception of MgO and suggests that this component accounts for hydration of the tholeiitic rocks. Both the first and second components (59% of the total variation) reflect most of the compositional variation of the tholeiitic rocks that can be explained by magmatic processes.

A graphical description of the relationships between the samples and the major element oxides for the first and second components is shown in Figure 28.13. Each rock type (THFE, THMG, THBA, THAN, THDA, THRH) has been individually contoured so that the relative densities and overlap between the groups can be observed. The relationships of the samples and variables indicate that iron-rich tholeiitic basalts are associated with $FeO^T$ and magnesium-rich tholeiitic basalts are associated with MgO and LOI. The fractionated tholeiitic rocks, andesite, dacite and rhyolite show association with relative $Na_2O$ and $K_2O$ enrichment. The normal tholeiitic basaltic rocks are located in the centre of the sample cloud, indicating little relative enrichment with respect to the other rock types. $TiO_2$-enriched iron-rich tholeiitic basalts are indicated as a group of samples that plot along the positive side of the PC2 axis. This diagram illustrates the fractionated characteristics of the tholeiitic rocks as well as the distinction between the major tholeiitic basalts. As a variation diagram it is more effective than the AFM or AFTiM diagrams since it describes the variation of the major element oxides simultaneously.

## CALC-ALKALIC REFERENCE GROUP

Logcontrast principal components analysis was applied to the calc-alkalic reference group and the results are shown in Table 28.7. The first 5 components account for more than 89% of the total data variation. The relative contributions of the variables for the first component show that with the exception of $P_2O_5$ and LOI, a significant proportion of the variation of the major element oxides is accounted for in the first component (53% of the data variation). The second component (15% of the variation) accounts for the variation of $P_2O_5$, CaO, $Al_2O_3$ and LOI. The third component (11% of the variation) accounts for LOI and $P_2O_5$. Together, the first and second components account for 68% of the variation in the data related to primary compositional variation.

The R-mode loadings of the first component show an inverse association of $FeO^T$, MgO, MnO, $TiO_2$ and CaO with $SiO_2$, $Na_2O$, $K_2O$ and $Al_2O_3$ and reflect the primary magmatic fractionation pattern of the calc-alkalic suite. Figure 28.14 shows the projection of samples and variables onto the PC1-PC2 axes. Each rock type (CABA, CAAN, CADA, CARH) has been individually contoured so that the relative densities and overlap between the groups can be observed. The diagram clearly shows the transitional association of the calc-alkalic basalts to andesites, and dacite to rhyolite. The mafic rocks are associated with MgO, $FeO^T$, MnO, $TiO_2$ and CaO at the negative end of the PC1 axis. The fractionated felsic rocks are associated with $K_2O$, $Na_2O$ and $SiO_2$ along the positive side of the PC1 axis. Variation along the PC2 axis shows relative enrichment in LOI and $P_2O_5$ along the negative part of the PC2 axis. CaO and $Al_2O_3$ enrichment, probably associated with plagioclase-enriched calc-alkalic rocks, occurs along the positive side of the PC2 axis.

## PRINCIPAL COMPONENTS ANALYSIS APPLIED TO NORMATIVE MINERALS TO ASSESS COMPOSITIONAL VARIATION

The use of geochemistry in assessing the variation of volcanic rocks is based upon fixed compositions of minerals. It is the relative abundance of minerals of fixed compositions that gives rise to the chemical variation. When normative minerals are used to assess compositional variation, then a model is implied about the association of the major element oxides. The use of normative minerals is

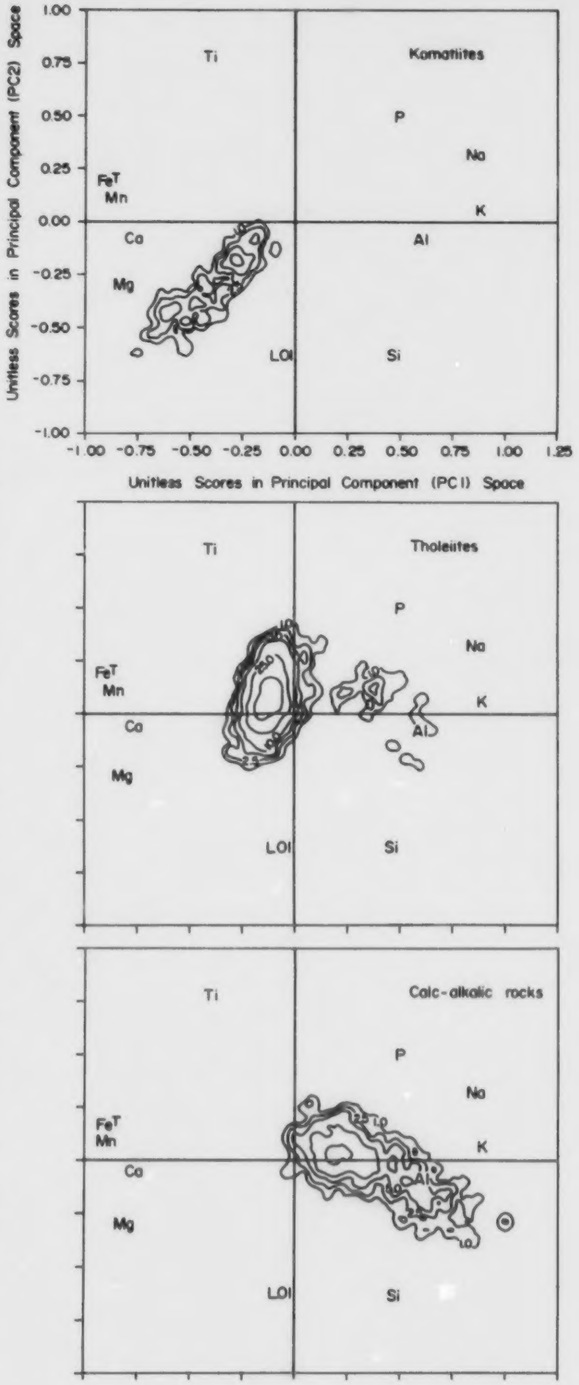

**Figure 28.11.** Multiplicative Logratio Principal Components Analysis PC1-PC2 PETROCH Reference Groups. Oxides are plotted as elements only. The relative positions of the elements indicate the relative relationships of the elements with respect to the magma clan reference groups: komatiites, tholeiites and calc-alkalic rocks. Each reference group is individually contoured. The contour lines represent densities per unit cell in the figure. Each reference group is contoured separately so that relative densities and overlap can be observed. The figure was divided into 100 cells along the x-axis, 80 cells along the y-axis. Contour intervals are: 1, 2.5, 5, 10, 25 and 50 points per square unit.

**Table 28.5.** Multiplicative logratio principal components analysis: PETROCH komatiitic reference group. [KOUM/KOBA] Samples: 155

| Component | Eigenvalue | %Trace | Σ%Trace |
|---|---|---|---|
| 1 | 5.5534 | 50.4855 | 50.4855 |
| 2 | 1.7420 | 15.8362 | 66.3217 |
| 3 | 1.2539 | 11.3991 | 77.7207 |
| 4 | 0.7634 | 6.9403 | 84.6611 |
| 5 | 0.5857 | 5.3243 | 89.9854 |
| 6 | 0.4542 | 4.1289 | 94.1142 |
| 7 | 0.3419 | 3.1084 | 97.2226 |
| 8 | 0.1997 | 1.8152 | 99.0378 |
| 9 | 0.0625 | 0.5681 | 99.6060 |
| 10 | 0.0433 | 0.3940 | 100.0000 |

| Component* | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| **Principal Component Loadings** | | | | | |
| Si | 0.8692 | 0.3787 | -0.0190 | -0.1057 | 0.0544 |
| Al | -0.6328 | 0.3070 | 0.1477 | 0.4823 | -0.3121 |
| $Fe^T$ | 0.8955 | 0.2841 | -0.0001 | 0.1416 | 0.0419 |
| Mg | 0.9553 | -0.1477 | -0.1153 | -0.0019 | 0.0964 |
| Ca | 0.1337 | 0.8317 | 0.0535 | -0.4313 | -0.2355 |
| Na | -0.8479 | 0.0278 | -0.0178 | -0.3329 | 0.2404 |
| K | -0.7390 | 0.0495 | -0.5143 | 0.1835 | -0.1202 |
| Ti | -0.2953 | 0.4894 | 0.5938 | 0.2888 | 0.4365 |
| P | 0.0747 | -0.5045 | 0.7703 | -0.1571 | -0.3116 |
| Mn | 0.7649 | 0.2657 | -0.0070 | 0.1884 | -0.2178 |
| LOI | 0.8497 | -0.3770 | -0.0694 | 0.1600 | 0.1069 |
| **Relative Contributions (%)** | | | | | |
| Si | 75.5578 | 14.3449 | 0.0362 | 1.1164 | 0.2965 |
| Al | 40.0447 | 9.4262 | 2.1809 | 23.2602 | 9.7423 |
| $Fe^T$ | 80.1893 | 8.0732 | 0.0000 | 2.0049 | 0.1752 |
| Mg | 91.2620 | 2.1804 | 1.3299 | 0.0004 | 0.9302 |
| Ca | 1.7870 | 69.1726 | 0.2863 | 18.6020 | 5.5469 |
| Na | 71.9007 | 0.0773 | 0.0318 | 11.0790 | 5.7807 |
| K | 54.6132 | 0.2447 | 26.4503 | 3.3663 | 1.4446 |
| Ti | 8.7185 | 23.9548 | 35.2582 | 8.3380 | 19.0568 |
| P | 0.5585 | 25.4552 | 59.3297 | 2.4666 | 9.7077 |
| Mn | 58.5030 | 7.0573 | 0.0049 | 3.5512 | 4.7434 |
| LOI | 72.2057 | 14.2113 | 0.4816 | 2.5588 | 1.1428 |

Major element oxides are denoted by their elements only.

* *Only the first 5 components are listed.*

also subject to the unit sum constraint and thus requires the use of logratio methods for any interpretation of the variation of the data. Barth-Niggli normative minerals were computed for the 3 reference suites. The normative mineral compositions were then subjected to logcontrast principal components analysis in order to examine the modal mineral variation of the suites.

Table 28.8 and Figure 28.15 show the results of logcontrast principal components analysis applied to the normative minerals of the tholeiitic reference suite. The first 5 components account for 76% of the data variation. The first 3 components are of the same order of magnitude, 25%, 24% and 11%, respectively.

The relative contributions of the first component show that the variation is accounted for by enstatite, ferrosilite, fayalite and forsterite. The relative contributions of the second component indicate that most of the variation is dominated by diopside, quartz, apatite, orthoclase, albite, anorthite, fayalite and forsterite.

Figure 28.15 shows the sample and normative mineral scores plotted onto the first 2 component axes. Each rock type (THFE, THMG, THBA, THAN, THDA, THRH) has been individually contoured so that the relative densities and overlap between groups can be observed. The second component (24% of the data variation) accounts for most of the forsterite and fayalite variation with an inverse association with enstatite and ferrosilite. This also reflects the distinction between undersaturated (olivine normative) and saturated (quartz normative) tholeiitic basalts. Note that many of the basalts cluster near the centre of the plot, indicating neither olivine nor quartz enrichment. These samples can be considered to represent the tholeiitic rocks that lie between the critical plane of undersaturation (diopside-forsterite-albite) and the plane defining normative quartz in the norm (diopside-enstatite-albite) (Cox et al. 1979, p.229).

The third component (11% of the variation) is dominated by pyrite, with lesser contributions from

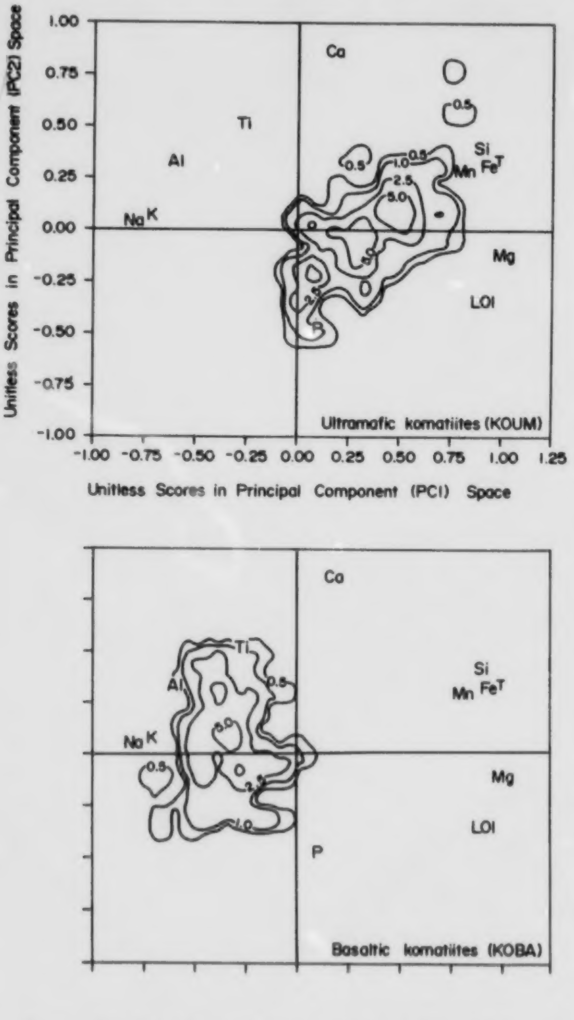

**Figure 28.12.** Multiplicative Logratio Principal Components Analysis PC1-PC2 PETROCH: Komatiitic Reference Groups. Oxides are plotted as elements only. The relative positions of the elements indicate the relative relationships of the elements with respect to the komatiitic reference groups. Each reference group is individually contoured. The contour lines represent densities per unit cell in the figure. The figure was divided into 45 cells along the x-axis, 40 cells along the y-axis. Each reference group is contoured separately so that relative densities and overlap can be observed.

magnetite and albite, and suggests that pyrite variation is more significant than variation of any of the other normative minerals.

The separation between the 3 mafic basaltic rock types is more distinct in Figure 28.13, where the PCA has been carried out on the major element oxides. This separation is not as apparent in Figure 28.15, where the 3 rock types are mixed together in the region between the critical plane of undersaturation and the plane defining normative quartz. This is due to the inverse relationship of forsterite with fayalite and enstatite with ferrosilite. Since both the olivine and pyroxene minerals contain iron and magnesium, and all 3 basaltic rock types contain varying proportions of the olivine and pyroxene minerals, the distinction between iron-rich and magnesium-rich basaltic rocks is masked.

The application of logcontrast principal components analysis has shown, in the case of normative minerals, that characteristic features of the igneous systems can be extracted. In this particular example, the critical zones of undersaturation and quartz saturation show up distinctly in the first 2 principal components.

# CANONICAL VARIATE ANALYSIS

Statistical procedures, including canonical variate analysis (multiple discriminant functions analysis), have been used by Pearce (1976), Chayes (1966, 1979, 1980), LeMaitre (1968, 1976a, 1976b, 1976c, 1982) and Muir (1979) to classify volcanic rocks. The most common statistical approach of classifying volcanic rocks is using canonical variate analysis (CVA) from which reference populations are created with distinct geochemical attributes. CVA and the associated linear discriminant functions test the statistical uniqueness of the reference groups and indicate whether the distinctions between the groups are significant. A description of the method can be found in LeMaitre (1982), Cooley and Lohnes (1971) and Srivastava and Carter (1983), with subsequent modifications as outlined by Aitchison (1986, p.202–206). Appendix 2 provides a brief description of the method.

The existence of singular covariance matrices is a particular problem when the multiplicative logratio is used in this type of analysis. One way of overcoming this problem is to apply a generalized inverse routine. Alternatively, by using additive logratio transforms as discussed previously, the problem of singularity can be avoided, with no loss of information with respect to the variability of the compositions. Additive logratios were calculated for each reference group and subsequent statistical procedures were carried out on the transformed data.

Canonical variate analysis assumes that the covariances of the groups (obtained from logratio transformations) are equal. A multiple analysis of variance procedure (MANOVA) was carried out on the 3 reference groups. The results indicated that the covariances of the 3 groups are not equal. Unequal covariance matrices may cause the discriminant function surface to be non-linear and thus the distinction between the groups may be difficult to interpret (LeMaitre 1982, p.142). Nonetheless, the results of the CVA indicate that the groups are sufficiently distinct.

Canonical variate analysis was applied to the 3 magma clan reference groups and the 12 individual reference groups that define the rock types. The additive logratio was used for the transformation and LOI was chosen as the divisor.

Canonical variate analysis was applied to the following groups of data: the 3 major magma clan reference groups (komatiites, tholeiites and calc-alkalic rocks); komatiitic rock types (KOUM and KOBA); tholeiitic rock types (THFE, THMG, THBA, THAN, THDA and THRH); and calc-alkalic rock types (CABA, CAAN, CADA and CARH).

**Table 28.6.** Multiplicative logratio principal components analysis: PETROCH tholeiitic reference group. Transforms [THFE, THMG, THBA, THAN, THDA, THRH] Samples: 1112

| Component | Eigenvalue | %Trace | Σ%Trace |
|---|---|---|---|
| 1 | 3.9871 | 36.2461 | 36.2461 |
| 2 | 2.5478 | 23.1617 | 59.4078 |
| 3 | 1.3398 | 12.1797 | 71.5875 |
| 4 | 1.0671 | 9.7011 | 81.2885 |
| 5 | 0.7350 | 6.6815 | 87.9700 |
| 6 | 0.4755 | 4.3231 | 92.2931 |
| 7 | 0.4041 | 3.6735 | 95.9666 |
| 8 | 0.2148 | 1.9525 | 97.9192 |
| 9 | 0.1184 | 1.0762 | 98.9954 |
| 10 | 0.1105 | 1.0046 | 100.0000 |

| Component* | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| **Principal Component Loadings** | | | | | |
| Si | 0.1128 | -0.8386 | -0.2509 | 0.2901 | -0.0501 |
| Al | 0.4531 | -0.7952 | -0.0901 | 0.2030 | -0.1736 |
| $Fe^T$ | 0.8053 | 0.3600 | -0.2733 | 0.1412 | 0.1192 |
| Mg | 0.8421 | -0.0284 | 0.2023 | -0.2416 | -0.1197 |
| Ca | 0.8851 | -0.1714 | -0.0201 | -0.2007 | -0.1983 |
| Na | -0.4616 | -0.5326 | -0.2770 | 0.3413 | 0.2795 |
| K | -0.7042 | -0.1519 | -0.2306 | -0.5901 | 0.2248 |
| Ti | 0.0868 | 0.7124 | -0.3890 | 0.4122 | 0.1653 |
| P | -0.5753 | 0.4297 | -0.1478 | 0.2339 | -0.6070 |
| Mn | 0.7600 | 0.2298 | -0.2047 | 0.0046 | 0.2625 |
| LOI | -0.0548 | 0.0264 | 0.8987 | 0.3659 | 0.2027 |
| **Relative Contributions (%)** | | | | | |
| Si | 1.2718 | 70.3326 | 6.2959 | 8.4178 | 0.2513 |
| Al | 20.5297 | 63.2353 | 0.8124 | 4.1210 | 3.0139 |
| $Fe^T$ | 64.8520 | 12.9575 | 7.4714 | 1.9935 | 1.4203 |
| Mg | 70.9108 | 0.0808 | 4.0915 | 5.8362 | 1.4331 |
| Ca | 78.3330 | 2.9382 | 0.0406 | 4.0261 | 3.9328 |
| Na | 21.3065 | 28.3638 | 7.6741 | 11.6507 | 7.8112 |
| K | 49.5877 | 2.3072 | 5.3195 | 34.8192 | 5.0546 |
| Ti | 0.7532 | 50.7521 | 15.1317 | 16.9881 | 2.7319 |
| P | 33.0964 | 18.4612 | 2.1853 | 5.4699 | 36.8470 |
| Mn | 57.7659 | 5.2805 | 4.1899 | 0.0021 | 6.8926 |
| LOI | 0.3003 | 0.0695 | 80.7641 | 13.3870 | 4.1080 |

Major element oxides are denoted by their elements only.

* *Only the first 5 components are listed.*

## Canonical Variate Analysis: Komatiite, Tholeiite and Calc-Alkalic Magma Clan Reference Groups

Table 28.9 shows the results of canonical variate analysis (CVA) for the 3 magma clan reference groups (KOM, THOL and CALC). The Wilk's Λ value of 0.0796 indicates the overall distinction between the 3 groups is significant. This is also indicated by the high F-ratio of the 3 groups (259.04).

The first canonical r value of 0.905 indicates that the distinction between the groups is significant. Similarly, the second canonical r has a value of 0.778, which indicates that it is almost as effective at distinguishing the groups but not as good as the first function. The first discriminant function accounts for 61% of the information, while the second function accounts for 39% of the information. Both $\chi^2$ tests give large values and indicate that the 2 discriminant functions are significant at the 99% confidence level.

The canonical variate scores of the 2 discriminant functions indicate the relative relationships of the major element oxides with respect to each other. The canonical variate scores indicate that $Na_2O$, $K_2O$ and MgO are the extreme scores and represent the extremes in compositional variation. The centroid of each group is listed in Table 28.9. The relationships of the coefficients of the first canonical variate are best described graphically. Figures 28.16a and 28.16b show the relationships of the major element oxides and the linear discriminant function scores of the samples for the first 2 canonical variates. Figure 28.16a displays the relative relationships of the major element oxides and when compared with the discriminant scores of the samples in Figure 28.16b. Each magma clan (KOM, THOL, CALC) has been individually contoured so that the relative densities and overlap can be seen. Figure 28.16b shows that

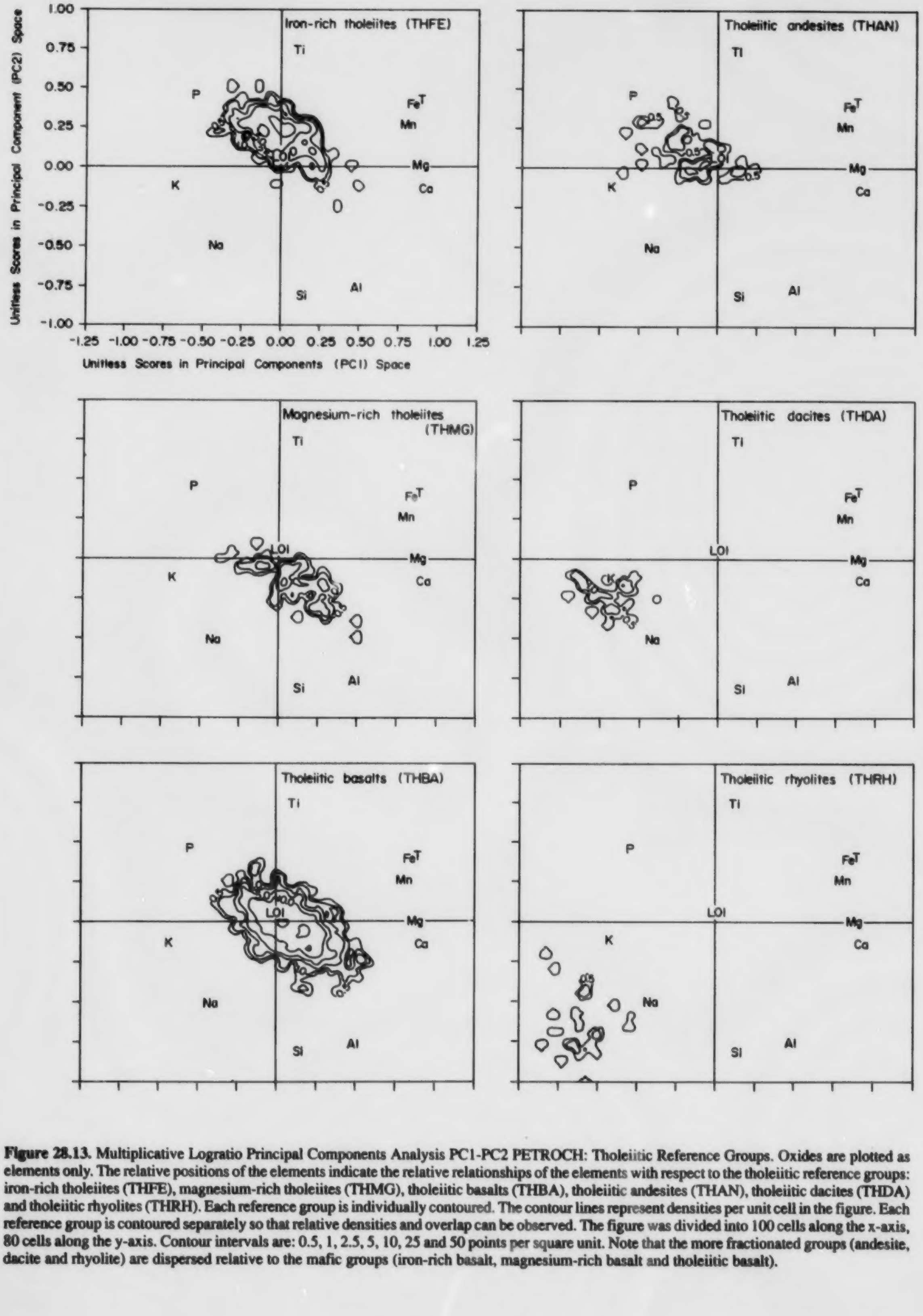

**Figure 28.13.** Multiplicative Logratio Principal Components Analysis PC1-PC2 PETROCH: Tholeiitic Reference Groups. Oxides are plotted as elements only. The relative positions of the elements indicate the relative relationships of the elements with respect to the tholeiitic reference groups: iron-rich tholeiites (THFE), magnesium-rich tholeiites (THMG), tholeiitic basalts (THBA), tholeiitic andesites (THAN), tholeiitic dacites (THDA) and tholeiitic rhyolites (THRH). Each reference group is individually contoured. The contour lines represent densities per unit cell in the figure. Each reference group is contoured separately so that relative densities and overlap can be observed. The figure was divided into 100 cells along the x-axis, 80 cells along the y-axis. Contour intervals are: 0.5, 1, 2.5, 5, 10, 25 and 50 points per square unit. Note that the more fractionated groups (andesite, dacite and rhyolite) are dispersed relative to the mafic groups (iron-rich basalt, magnesium-rich basalt and tholeiitic basalt).

**Table 28.7. Multiplicative logratio principal components analysis: PETROCH calc-alkalic reference group. [CABA, CAAN, CADA, CARH] Samples: 578**

| Component | Eigenvalue | %Trace | Σ%Trace |
|---|---|---|---|
| 1 | 5.8093 | 52.8121 | 52.8121 |
| 2 | 1.6474 | 14.9762 | 67.7883 |
| 3 | 1.1940 | 10.8550 | 78.6433 |
| 4 | 0.6571 | 5.9734 | 84.6167 |
| 5 | 0.5448 | 4.9530 | 89.5696 |
| 6 | 0.4673 | 4.2485 | 93.8182 |
| 7 | 0.3762 | 3.4202 | 97.2384 |
| 8 | 0.1514 | 1.3761 | 98.6145 |
| 9 | 0.0857 | 0.7793 | 99.3938 |
| 10 | 0.0667 | 0.6062 | 100.0000 |

| Component* | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| **Principal Component Loadings** | | | | | |
| Si | 0.9073 | 0.3321 | -0.0121 | 0.0747 | -0.0313 |
| Al | 0.8027 | 0.5193 | -0.0385 | 0.1059 | -0.1561 |
| $Fe^T$ | -0.9057 | 0.1706 | -0.0405 | -0.0736 | 0.0281 |
| Mg | -0.8490 | -0.0353 | -0.0198 | -0.1800 | -0.2358 |
| Ca | -0.5776 | 0.6076 | -0.0070 | -0.0858 | -0.3834 |
| Na | 0.8792 | 0.2150 | -0.0191 | 0.1913 | -0.0115 |
| K | 0.8204 | -0.0151 | -0.0852 | -0.3735 | 0.2773 |
| Ti | -0.6730 | 0.2421 | -0.0696 | 0.5958 | 0.2404 |
| P | -0.0071 | -0.6515 | -0.6947 | 0.1480 | -0.1378 |
| Mn | -0.7496 | 0.2961 | 0.0607 | -0.1426 | 0.3875 |
| LOI | 0.0779 | -0.5007 | 0.8316 | 0.1474 | -0.1102 |
| **Relative Contributions (%)** | | | | | |
| Si | 82.3191 | 11.0284 | 0.0147 | 0.5582 | 0.0978 |
| Al | 64.4407 | 26.9636 | 0.1486 | 1.1223 | 2.4365 |
| $Fe^T$ | 82.0290 | 2.9102 | 0.1638 | 0.5424 | 0.0792 |
| Mg | 72.0771 | 0.1244 | 0.0393 | 3.2392 | 5.5610 |
| Ca | 33.3662 | 36.9135 | 0.0049 | 0.7365 | 14.6972 |
| Na | 77.3008 | 4.6238 | 0.0365 | 3.6598 | 0.0131 |
| K | 67.3075 | 0.0229 | 0.7259 | 13.9506 | 7.6881 |
| Ti | 45.2955 | 5.8602 | 0.4838 | 35.5001 | 5.7812 |
| P | 0.0051 | 42.4480 | 48.2605 | 2.1912 | 1.8975 |
| Mn | 56.1850 | 8.7694 | 0.3679 | 2.0336 | 15.0156 |
| LOI | 0.6073 | 25.0740 | 69.1590 | 2.1733 | 1.2153 |

Major element oxides are denoted by their elements only.

* *Only the first 5 components are listed.*

the komatiitic rocks plot at the positive side of the CV1 axis, while the tholeiitic rocks plot near the origin of the diagram. The calc-alkalic rocks plot in the quadrant of negative CV1 and positive CV2. There is a degree of overlap between the 3 magma clan reference groups, as shown by the overlapping boundaries, however, the canonical coordinates of the 3 main magma clans are clearly distinct in Figure 28.16b. The centroid of each group, listed in Table 28.9, also indicates that the magma clans are clearly distinct from each other.

## Komatiitic Reference Group

Canonical variate analysis was applied to the 2 komatiitic rock types (KOUM and KOBA). Because there are only 2 reference groups for this analysis, there is only 1 discriminant function. The results of the analysis are shown in Table 28.10. The value of Wilk's Λ of 0.1349 and the F-ratio for overall discrimination is 57.73 and indicates that the discrimination between the 2 groups is significant. The communalities of the variables for the discriminant function indicate that $Na_2O$, $Al_2O_3$ and $TiO_2$ are the most significant variables for discriminating between the 2 groups. MgO is an important discriminator between the 3 main magma clans, however, it is worth noting that the communality of MgO is low and indicates that MgO is not an important discriminator within the komatiite reference groups. The significance of $Na_2O$ may be due to effects of alteration, while the significance of $Al_2O_3$ and $TiO_2$ are distinctly associated with the basaltic komatiite reference group. The canonical scores of the variables show that MgO is the lowest score. The canonical coordinate of the ultramafic komatiite centroid is negative and is associated with the low canonical score of MgO. Figure 28.17a shows the plot of the canonical scores of the variables along the first canonical axis. The diagram displays an inverse relationship of MgO with $Na_2O$, $Al_2O_3$ and $TiO_2$, which reflects the main compositional differences between the 2 komatiitic rock types. Figure 28.17b shows the canonical variate scores of the samples. The figure shows that the 2 groups are clearly distinct. These samples are expressed as histograms in Figure 28.17c which shows the density distribution of the samples.

## Tholeiitic Reference Group

Canonical variate analysis was applied to the 6 tholeiitic reference groups (THFE, THMG, THBA, THAN, THDA and THRH). Table 28.11 shows the results of the analysis. A value of Wilk's $\Lambda$ of 0.0143 and the F-ratio of 80.30 indicates that the distinction between the groups is very significant. The first 2 canonical r values indicate that the first 2 linear discriminant functions are significant, however, the first function is by far the most significant (trace = 75%). The relative relationships of the major elements from the canonical variate scores indicates that $K_2O$, $Na_2O$, $P_2O_5$ and $SiO_2$ are inversely related to MgO, CaO, $FeO^T$, MnO and $TiO_2$. Also, for the second discriminant function, the canonical score of MgO is inversely related to the scores of $FeO^T$ and $TiO_2$. The communalities of the variables show that MgO and $FeO^T$ are the 2 most significant variables for discrimination between the tholeiitic rock types.

Figure 28.18a shows the canonical scores of the major element oxides plotted onto the first 2 canonical axes. The

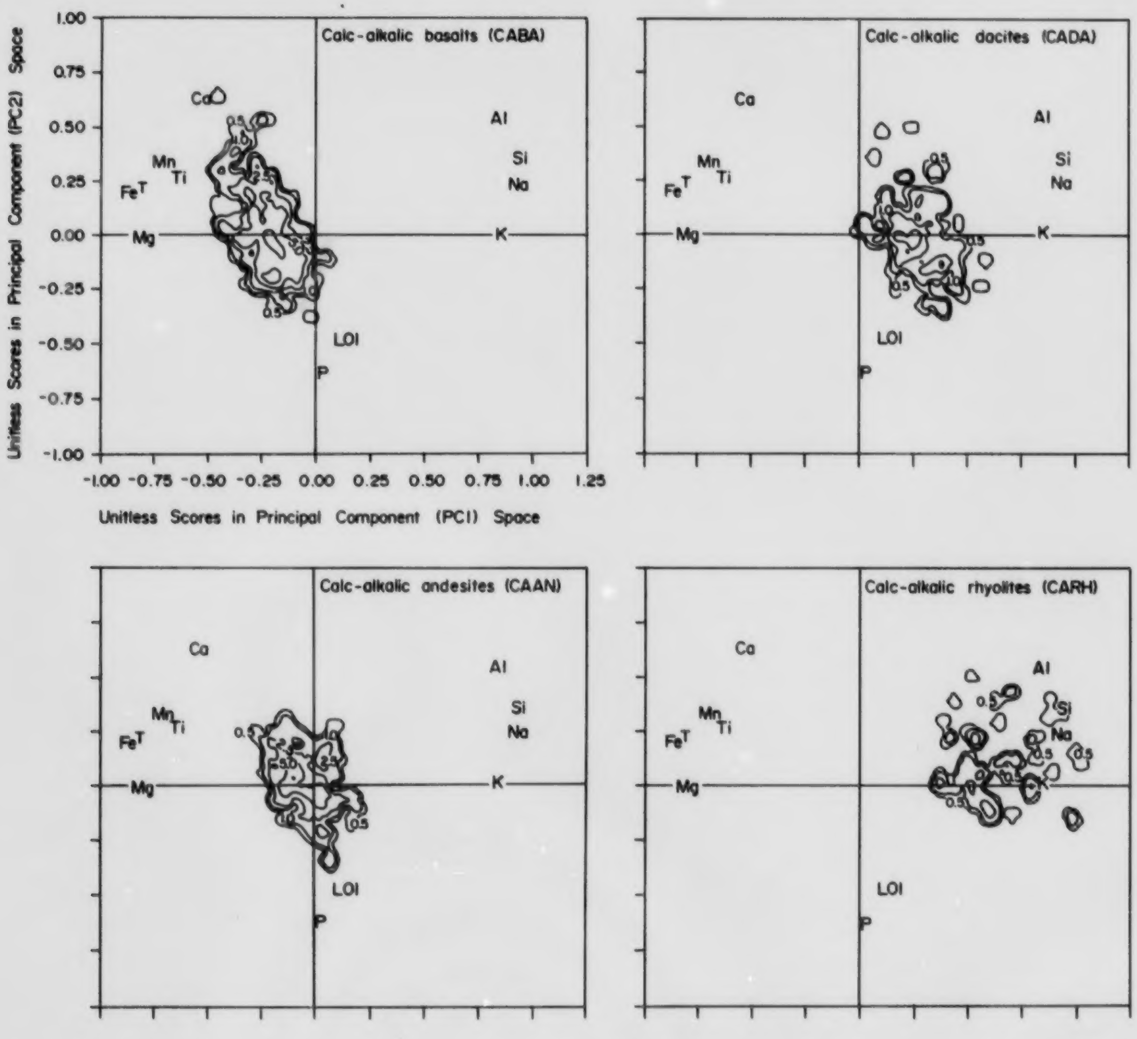

**Figure 28.14.** Multiplicative Logratio Principal Components Analysis PC1-PC2 PETROCH: Calc-Alkalic Reference Groups. Oxides are plotted as elements only. The relative positions of the elements indicate the relative relationships of the elements with respect to the calc-alkalic reference groups: calc-alkalic basalts (CABA), calc-alkalic andesites (CAAN), calc-alkalic dacites (CADA) and calc-alkalic rhyolites (CARH). Each reference group is individually contoured. The contour lines represent densities per unit cell in the figure. Each reference group is contoured separately so that the relative densities and overlap can be observed. The figure was divided into 100 cells along the x-axis, 80 cells along the y-axis. Contour intervals are: 0.5, 1, 2 and 5 points per square unit. Note that the more fractionated groups (dacite and rhyolite) are dispersed relative to the mafic groups (basalt and andesite).

**Table 28.8.** Multiplicative logratio principal components analysis: PETROCH tholeiitic reference groups—normative minerals. [THFE, THMG, THBA, THAN, THDA, THRH] Samples: 1112

| **Component** | **Eigenvalue** | **% Trace** | **Σ% Trace** |
|---|---|---|---|
| 1 | 3.5136 | 25.0968 | 25.0968 |
| 2 | 3.3799 | 24.1423 | 49.2391 |
| 3 | 1.5607 | 11.1478 | 60.3868 |
| 4 | 1.2270 | 8.7640 | 69.1509 |
| 5 | 1.0258 | 7.3270 | 76.4779 |
| 6 | 0.8899 | 6.3567 | 82.8346 |
| 7 | 0.7784 | 5.5600 | 88.3946 |
| 8 | 0.5477 | 3.9121 | 92.3067 |
| 9 | 0.5262 | 3.7587 | 96.0654 |
| 10 | 0.4090 | 2.9211 | 98.9865 |
| 11 | 0.1116 | 0.7972 | 99.7837 |
| 12 | 0.0246 | 0.1759 | 99.9596 |
| 13 | 0.0057 | 0.0404 | 100.0000 |

| **Component*** | **1** | **2** | **3** | **4** | **5** |
|---|---|---|---|---|---|
| **Principal Component Loadings** | | | | | |
| Aptt | 0.1369 | 0.6078 | -0.2742 | -0.2689 | 0.3405 |
| Pyrt | 0.0874 | 0.1208 | -0.7683 | 0.3397 | -0.1623 |
| Ilmn | -0.3362 | 0.4070 | -0.1841 | -0.3189 | 0.6184 |
| Orth | 0.4179 | 0.4523 | 0.2409 | 0.2887 | -0.1011 |
| Albt | 0.3475 | 0.3526 | 0.4494 | -0.3260 | 0.1070 |
| Anrt | -0.5663 | -0.3983 | 0.2666 | -0.3696 | -0.1928 |
| Mgnt | -0.3610 | 0.1749 | 0.6054 | 0.2585 | 0.2182 |
| Enst | -0.8593 | -0.0118 | 0.0143 | -0.2129 | -0.3424 |
| Fers | -0.7820 | 0.2799 | -0.2127 | -0.3030 | -0.2028 |
| Diop | -0.3529 | -0.8377 | 0.1314 | 0.2588 | 0.1966 |
| Qrtz | 0.0595 | 0.6336 | 0.3038 | 0.2332 | -0.1825 |
| Fors | 0.7149 | -0.5665 | 0.0476 | -0.3184 | -0.0893 |
| Faya | 0.7364 | -0.5153 | -0.0362 | -0.3616 | -0.0500 |
| Hedn | -0.3104 | -0.7375 | -0.0716 | 0.2322 | 0.3923 |
| **Relative Contributions (%)** | | | | | |
| Aptt | 1.8746 | 36.9460 | 7.5191 | 7.2289 | 11.5908 |
| Pyrt | 0.7631 | 1.4581 | 59.0246 | 11.5426 | 2.6333 |
| Ilmn | 11.3044 | 16.5675 | 3.3884 | 10.1699 | 38.2383 |
| Orth | 17.4640 | 20.4574 | 5.8034 | 8.3339 | 1.0214 |
| Albt | 12.0769 | 12.4351 | 20.2001 | 10.6272 | 1.1451 |
| Anrt | 32.0687 | 15.8667 | 7.1060 | 13.6601 | 3.7189 |
| Mgnt | 13.0333 | 3.0589 | 36.6540 | 6.6824 | 4.7592 |
| Enst | 73.8419 | 0.0140 | 0.0204 | 4.5339 | 11.7219 |
| Fers | 61.1457 | 7.8328 | 4.5256 | 9.1838 | 4.1119 |
| Diop | 12.4539 | 70.1664 | 1.7261 | 6.6967 | 3.8651 |
| Qrtz | 0.3545 | 40.1450 | 9.2310 | 5.4393 | 3.3307 |
| Fors | 51.1096 | 32.0891 | 0.2266 | 10.1359 | 0.7976 |
| Faya | 54.2304 | 26.5573 | 0.1307 | 13.0727 | 0.2500 |
| Hedn | 9.6342 | 54.3976 | 0.5126 | 5.3894 | 15.3934 |

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* *Only the first 5 components are listed.*

| | | | | | |
|---|---|---|---|---|---|
| Albt | Albite | Faya | Fayalite | Mgnt | Magnetite |
| Anrt | Anorthosite | Fers | Ferrosilite | Orth | Orthosite |
| Aptt | Apatite | Fors | Forsterite | Pyrt | Pyrite |
| Diop | Diopside | Hedn | Hedenbergite | Qrtz | Quartz |
| Enst | Enstatite | Ilmn | Ilmenite | | |

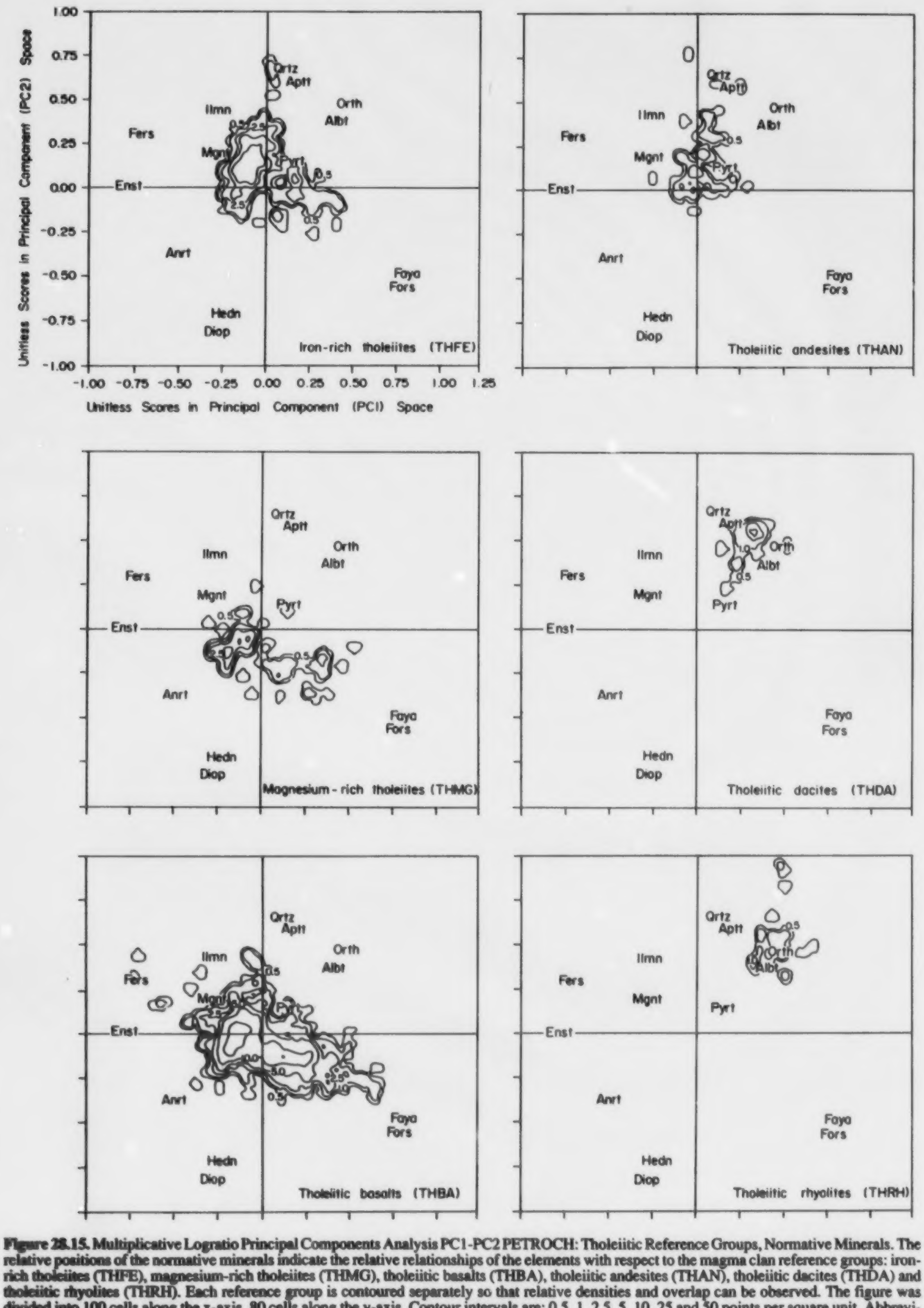

**Figure 28.15.** Multiplicative Logratio Principal Components Analysis PC1-PC2 PETROCH: Tholeiitic Reference Groups, Normative Minerals. The relative positions of the normative minerals indicate the relative relationships of the elements with respect to the magma clan reference groups: iron-rich tholeiites (THFE), magnesium-rich tholeiites (THMG), tholeiitic basalts (THBA), tholeiitic andesites (THAN), tholeiitic dacites (THDA) and tholeiitic rhyolites (THRH). Each reference group is contoured separately so that relative densities and overlap can be observed. The figure was divided into 100 cells along the x-axis, 80 cells along the y-axis. Contour intervals are: 0.5, 1, 2.5, 5, 10, 25 and 50 points per square unit. Abbreviations: Albt - Albite; Anrt - Anorthosite; Aptt - Apatite; Diop - Diopside; Enst - Enstatite; Faya - Fayalite; Fers - Ferrosilite; Fors - Forsterite; Hedn - Hedenbergite; Ilmn - Ilmenite; Mgnt - Magnetite; Orth - Orthosite; Pyrt - Pyrite; Qrtz - Quartz.

relationships of the elements can be related to the canonical scores of the samples in Figure 28.18b, which shows the canonical scores of the samples from the 6 tholeiitic reference groups. As in other figures, each rock type (THFE, THMG, THBA, THAN, THDA, THRH) has been individually contoured so that the relative densities and overlap can be observed. The distinctions between the groups are clearly shown. The iron-rich tholeiitic rocks plot along the negative side of the CV2 axis and correspond to the $Fe^T$- and titanium-enriched area of Figure 28.18a. The magnesium- rich tholeiitic basalts plot along the positive side of the CV2 axis and correspond to the magnesium-enriched area in Figure 28.18a. The normal tholeiitic basalts plot between the iron-rich and magnesium-rich basalts with minimal overlap. Since the boundaries between these 3 basaltic rock types are somewhat arbitrary, a certain degree of overlap is expected. The fractionated tholeiitic rocks, andesite, dacite and rhyolite, plot along the positive side of the CV1 axis. The increased dispersion of these groups, particularly the rhyolitic group, is a problem. This was primarily due to the difficulty in obtaining an adequate representation of these reference groups.

The first 2 discriminant functions account for more than 95% of the information. The first function accentuates the differences between the basalts from the fractionated tholeiitic rocks, while the second function highlights the differences between the 3 basaltic groups.

## Calc-Alkalic Reference Group

The results of CVA were applied to the 4 calc-alkalic reference groups (CABA, CAAN, CADA and CARH). The value of Wilk's $\Lambda$ of 0.0725 and the F-ratio of 42.81 indicate that the distinction between these reference groups is significant (Table 28.12). The first linear discriminant

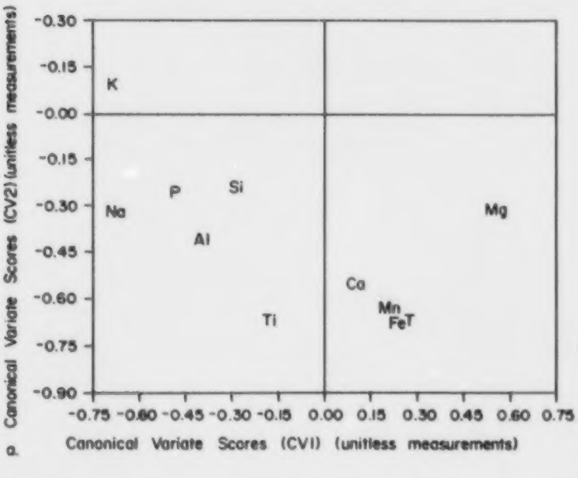

**Figure 28.16a.** Canonical Variate Analysis (CV1-CV2) PETROCH Reference Groups, Additive Logratio transformation. Plot of canonical scores of the major element oxides. Oxides are plotted as elements only. The relative positions of the elements indicate their relative relationships with respect to the magma clan reference groups.

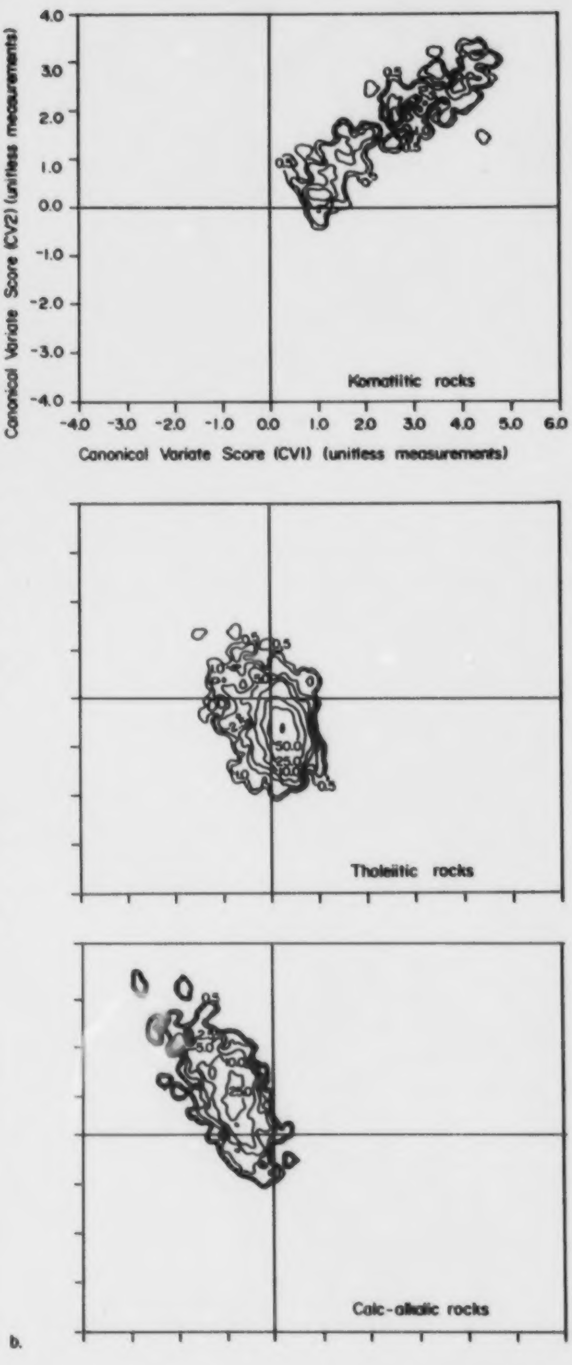

**Figure 28.16b.** Canonical Variate Analysis (CV1-CV2) PETROCH Magma Clan Reference Groups, Additive Logratio transformation. Plot of canonical scores of the reference group samples (1845 samples). The komatiites (KOM) plot in the positive quadrant associated with magnesium as shown in Figure 28.9. The tholeiitic (THOL) and calc-alkalic (CALC) rocks form 2 clouds of points that merge. Each reference group is contoured separately so that relative densities and overlap can be observed. The figure was divided into 100 cells along the x-axis, 80 cells along the y-axis. Contour intervals are: 0.5, 1, 2.5, 5, 10, 25 and 50 points per square unit.

**Table 28.9.** Canonical variate analysis—Three PETROCH magma clan reference groups: komatiites (KOM), tholeiites (THOL) and calc-alkalic rocks (CALC).

**Chi-Square Tests with Successive Roots Removed**

| Roots removed | Eigenvalue | $\chi^2$ | % Trace |
|---|---|---|---|
| 0 | 3.154 | 4650. | 60.93 |
| 1 | 2.023 | 2033. | 39.07 |

| | Discriminant Function Coefficients | | Canonical Scores | | Communalities for 2 Discriminant Functions |
|---|---|---|---|---|---|
| | DF1 | DF2 | CV1 | CV2 | |
| Si | 1.4433 | 1.7208 | -0.303 | -0.261 | 0.1597 |
| Al | -1.5757 | -1.0858 | -0.421 | -0.427 | 0.3588 |
| Fe$^T$ | 1.0631 | -1.9801 | 0.197 | -0.678 | 0.4986 |
| Mg | 0.0313 | 1.2720 | 0.518 | -0.325 | 0.3744 |
| Ca | -0.0136 | -0.2704 | 0.074 | -0.577 | 0.3390 |
| Na | -0.3884 | -0.1971 | -0.710 | -0.338 | 0.6181 |
| K | -0.1105 | 0.2569 | -0.705 | 0.079 | 0.5036 |
| Ti | -0.5396 | 0.1300 | -0.197 | -0.688 | 0.5119 |
| P | -0.1259 | 0.1523 | -0.501 | -0.270 | 0.3236 |
| Mn | 0.2558 | -0.1712 | 0.179 | -0.668 | 0.4775 |

**Canonical Coordinates of Group Centroids**

| | CV1 | CV2 |
|---|---|---|
| Komatiites (KOM) | 2.3743 | 1.5245 |
| Tholeiites (THOL) | 0.1527 | -0.6483 |
| Calc-alkalic rocks (CALC) | -0.9305 | 0.8385 |

Major element oxides are denoted by their elements only.

Wilk's $\Lambda$ = 0.0796

F-ratio for $h^2$ (overall discrimination) = 259.04

N.d.f.1 = 36 and N.d.f.2 = 3666

**Table 28.10.** Canonical variate analysis—PETROCH komatiitic reference groups (KOM): ultramafic komatiite (KOUM) and basaltic komatiite (KOBA).

**Chi-Square Tests With Successive Roots Removed**

| Roots removed | Eigenvalue | $\chi^2$ | % Trace |
|---|---|---|---|
| 0 | 6.414 | 297. | 100.00 |

| | Discriminant Function Coefficients | Canonical Scores | Communalities for 1 Discriminant Function |
|---|---|---|---|
| | DF1 | CV1 | |
| Si | 2.0219 | 0.800 | 0.6399 |
| Al | 0.3589 | 0.900 | 0.8091 |
| Fe$^T$ | -1.3124 | 0.780 | 0.6081 |
| Mg | -1.1557 | 0.464 | 0.2151 |
| Ca | -0.6061 | 0.792 | 0.6269 |
| Na | 0.2896 | 0.936 | 0.8764 |
| K | 0.0320 | 0.807 | 0.6515 |
| Ti | 0.6341 | 0.883 | 0.7793 |
| P | -0.0726 | 0.672 | 0.4514 |
| Mn | -0.0118 | 0.768 | 0.5903 |

**Canonical Coordinates of Group Centroids**

| | CV1 |
|---|---|
| Ultramafic komatiite (KOUM) | -0.8635 |
| Basaltic komatiite (KOBA) | 0.9954 |

Major element oxides are denoted by their elements only.

Wilk's $\Lambda$ = 0.1349

F-ratio for $h^2$ (overall discrimination) = 57.73

N.d.f.1 = 16 and N.d.f.2 = 144

function accounts for almost all of the distinctions between the groups (trace = 98.74%).

Figure 28.19a shows a plot of the first and second canonical scores of the variables. The major element oxides associated with the basaltic rocks, MgO, FeO$^T$, $TiO_2$, CaO and MnO, plot along the positive side of the CV1 axis, while the oxides associated with the fractionated felsic rocks, $K_2O$, $Na_2O$, $SiO_2$ and $Al_2O_3$ plot along the negative side of the CV1 axis. Figure 28.19b shows the canonical scores of the samples. The basalts and andesites (CABA, CAAN) plot along the positive side of the CV1 axis. The fractionated felsic rocks, dacites and rhyolites (CADA, CARH), plot along the negative side of the CV1 axis. Each rock type (CABA, CAAN, CADA, CARH) has been individually contoured so that the relative densities and overlap between the groups can be observed. As in the case for the tholeiitic rocks, the dispersion of the felsic reference groups is greater than the mafic reference groups. Again, this is mostly due to the inadequate number of samples available to construct reference groups with minimal alteration. Overlap exists between the basalts and andesites (CABA, CAAN) and between the dacites and rhyolites (CADA, CARH). This is expected as the boundaries between these rock types are transitional.

Canonical variate analysis has established that the magma clans and the rock types within each clan are statistically distinct. It follows that unknown samples should be tested for membership within none, one, or several of the reference groups. The use of allocation procedures provides 2 methods whereby unknown samples can be tested for similarity with the reference groups and, if possible, assign membership to one of the reference groups.

## APPLICATION OF ALLOCATION PROCEDURES TO PETROCH SAMPLES

The use of canonical variate analysis has provided a means of obtaining a statistical measure of the distinction of the reference groups. The analyses indicate that all of the magma clan and rock type reference groups are statistically distinct. A consequence of the establishment and testing of the reference groups is the ability to test unknown samples for possible membership into one of the groups. This type of procedure is known as allocation (also known as classification procedures).

Allocation procedures make it possible to predict the probability of an unknown sample belonging to a set of reference groups. The method is based on the assumption that the variables are the same between the reference groups and the unknown sample(s). A general reference to allocation procedures can be found in Cooley and Lohnes (1971), LeMaitre (1982), Campbell (1984) and Aitchison (1986).

Allocation methods have not been used extensively in geological classification analysis. Exceptions to this have been for purposes of geochemical exploration as documented by Smith et al. (1984) and Garrett (1989a). These contributions clearly explain the usefulness of the method.

Allocation procedures work on the basis of measuring the distance of a sample from each reference group centroid. By using the covariance estimate of the total population of samples used for the reference groups, the Mahalanobis distance can be computed between each unknown sample and the group centroids. LeMaitre (1982, p.171) provided a schematic concept of the allocation procedure. The covariance matrix contains the characteristics of the dispersion of the reference population. If the Mahalanobis distance of a sample, with respect to a particular reference group centroid, is within the dispersion matrix defined by the reference group covariances, then the probability of that sample belonging to that reference group is greater than zero. The closer a sample is to a group centroid, the higher the probability of membership, however, any sample with a probability greater than 0 is considered to be similar to the

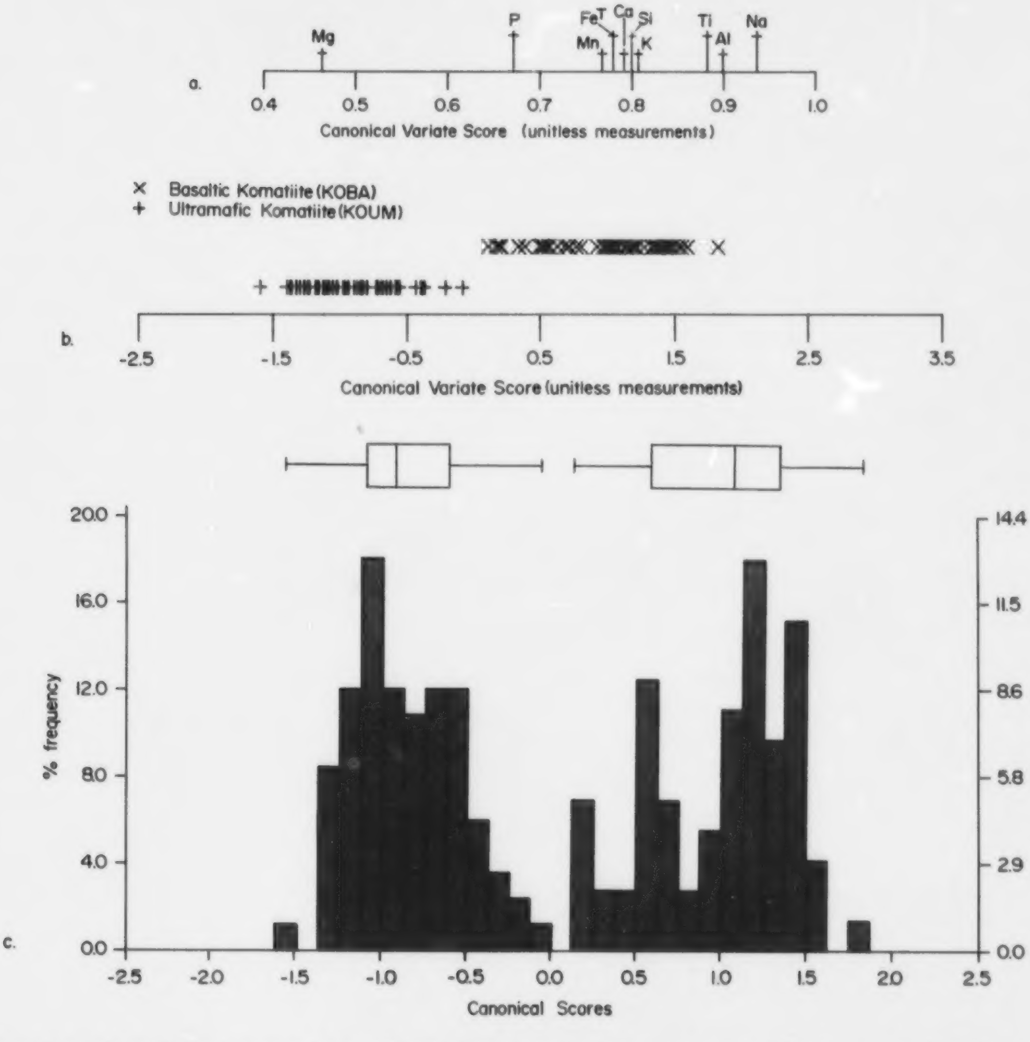

**Figure 28.17.** Canonical Variate Analysis (CV1-CV2) PETROCH Komatiitic Reference Groups, Additive Logratio transformation. **a)** Plot of canonical scores of the major element oxides. Oxides are plotted as elements only. The relative positions of the elements indicate their relative relationships with respect to the 2 komatiitic reference groups. MgO is distinctively separate from $Na_2O$, $TiO_2$ and $Al_2O_3$. **b)** Plot of canonical scores of the reference group samples (155 samples). The ultramafic komatiites plot along the negative side of the axis, while the basaltic komatiites plot along the positive side of the axis. **c)** Histogram plot of the canonical scores of the reference group samples (155 samples). The histogram for the ultramafic komatiites plots along the negative side of the axis, while the histogram for the basaltic komatiites plots along the positive side of the axis. The histograms assist in visualizing the relative density distributions of the reference groups.

**Table 28.11.** Canonical variate analysis—PETROCH tholeiitic reference groups (THOL): iron-rich tholeiite (THFE), magnesium-rich tholeiite (THMG), tholeiitic basalt (THBA), tholeiitic andesite (THAN), tholeiitic dacite (THDA) and tholeiitic rhyolite (THRH).

**Chi-Square Tests With Successive Roots Removed**

| Roots removed | Eigenvalue | $\chi^2$ | % Trace |
|---|---|---|---|
| 0 | 10.495 | 4689 | 75.45 |
| 1 | 2.883 | 1995 | 20.73 |
| 2 | 0.447 | 499 | 3.21 |
| 3 | 0.047 | 92 | 0.34 |
| 4 | 0.038 | 41 | 0.27 |

| | Discriminant Function Coefficients | | | | | Canonical Scores | | | | | Communalities for 5 Discriminant Functions |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | DF1 | DF2 | DF3 | DF4 | DF5 | CV1 | CV2 | CV3 | CV4 | CV5 | |
| Si | 1.8276 | 0.9012 | 8.2203 | 3.9327 | 1.0361 | 0.084 | -0.047 | 0.244 | -0.363 | -0.369 | 0.3367 |
| Al | 1.6777 | 1.9783 | -7.3727 | -4.6128 | -3.9154 | -0.005 | -0.008 | 0.144 | -0.396 | -0.400 | 0.3379 |
| $Fe^T$ | -0.8902 | -3.1952 | 4.4287 | -2.3812 | -3.7876 | -0.256 | -0.252 | 0.226 | -0.382 | -0.373 | 0.4649 |
| Mg | -1.3818 | 1.8525 | 1.1713 | -0.1580 | 1.9589 | -0.538 | 0.153 | 0.209 | -0.324 | -0.297 | 0.5498 |
| Ca | -0.7635 | -0.4464 | -1.5744 | 3.3269 | -0.1375 | -0.312 | 0.009 | 0.090 | -0.264 | -0.401 | 0.3364 |
| Na | -0.1631 | -0.1632 | -1.1144 | -0.7119 | -0.0843 | 0.198 | -0.105 | 0.140 | -0.453 | -0.282 | 0.3551 |
| K | 0.0323 | 0.0117 | 0.1580 | 0.2596 | 0.6327 | 0.364 | -0.115 | 0.210 | -0.191 | 0.138 | 0.2454 |
| Ti | -0.1191 | -0.7525 | -1.5010 | 2.9772 | 0.3797 | -0.121 | -0.469 | 0.130 | -0.249 | -0.255 | 0.3787 |
| P | -0.0100 | -0.0460 | 0.1485 | -0.7647 | 0.9829 | 0.174 | -0.379 | 0.136 | -0.472 | 0.138 | 0.4341 |
| Mn | -0.2052 | -0.1926 | -2.3745 | -2.7664 | 2.2577 | -0.222 | -0.210 | 0.108 | -0.433 | -0.299 | 0.3820 |

**Canonical Coordinates of Group Centroids**

| | CV1 | CV2 | CV3 | CV4 | CV5 |
|---|---|---|---|---|---|
| Iron-rich tholeiite (THFE) | -0.1494 | -1.3526 | 0.4300 | 0.0928 | 0.0324 |
| Magnesium-rich tholeiite (THMG) | -0.3848 | 1.6810 | 0.4641 | 0.4947 | 0.3774 |
| Tholeiitic basalt (THBA) | -0.2630 | 0.3644 | -0.0827 | -0.0965 | -0.0724 |
| Tholeiitic andesite (THAN) | 0.8760 | -0.8752 | -2.3032 | 0.3079 | 0.1384 |
| Tholeiitic dacite (THDA) | 3.2859 | -0.0123 | 0.2693 | -0.7715 | 0.8122 |
| Tholeiitic rhyolite (THRH) | 5.2867 | 0.8205 | 0.7539 | 0.4153 | -0.6058 |

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Major element oxides are denoted by their elements only.

Wilk's $\Lambda$ = 0.0143

F-ratio for $h^2$ (overall discrimination) = 80.30

N.d.f.1 = 96 and N.d.f.2 = 5006

reference group. The dispersion between groups can overlap, which results in some samples having probabilities of belonging to more than one reference group. Similarly, if a sample has a Mahalanobis distance outside the dispersion matrix, then it has a zero probability of belonging to any of the reference groups. A significant test is the determination of whether or not the groups have similar covariances. If they do not, then comparison of samples must be modified according to these differences (LeMaitre 1982).

A brief description of allocation procedures is provided in Appendix 3.

Additive logratios of the reference data sets and the altered rejected data were subjected to the allocation procedures. The reference data were classified for the purposes of verification. Tests of covariance similarities between the groups indicate that the covariances are not equal.

The results of the classification can best be illustrated by providing a list of representative sample compositions and the corresponding probabilities that are calculated with respect to the reference groups. Table 28.13a lists representative compositions of samples that were retrieved from the reference groups. The compositions are for the

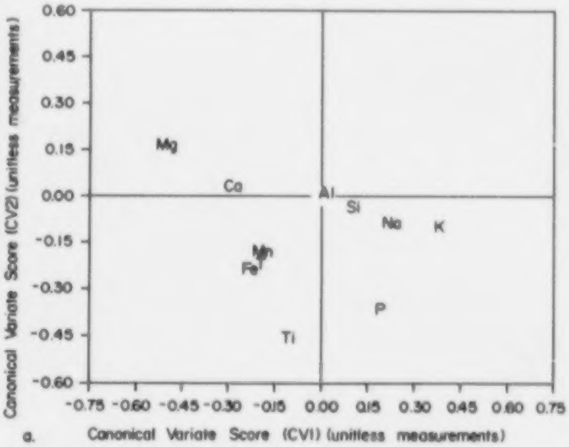

**Figure 28.18a.** Canonical Variate Analysis (CV1-CV2) PETROCH Tholeiitic Reference Groups, Additive Logratio transformation. Plot of canonical scores of the major element oxides. Oxides are plotted as elements only. The relative positions of the elements indicate their relative relationships with respect to the magma clan reference groups.

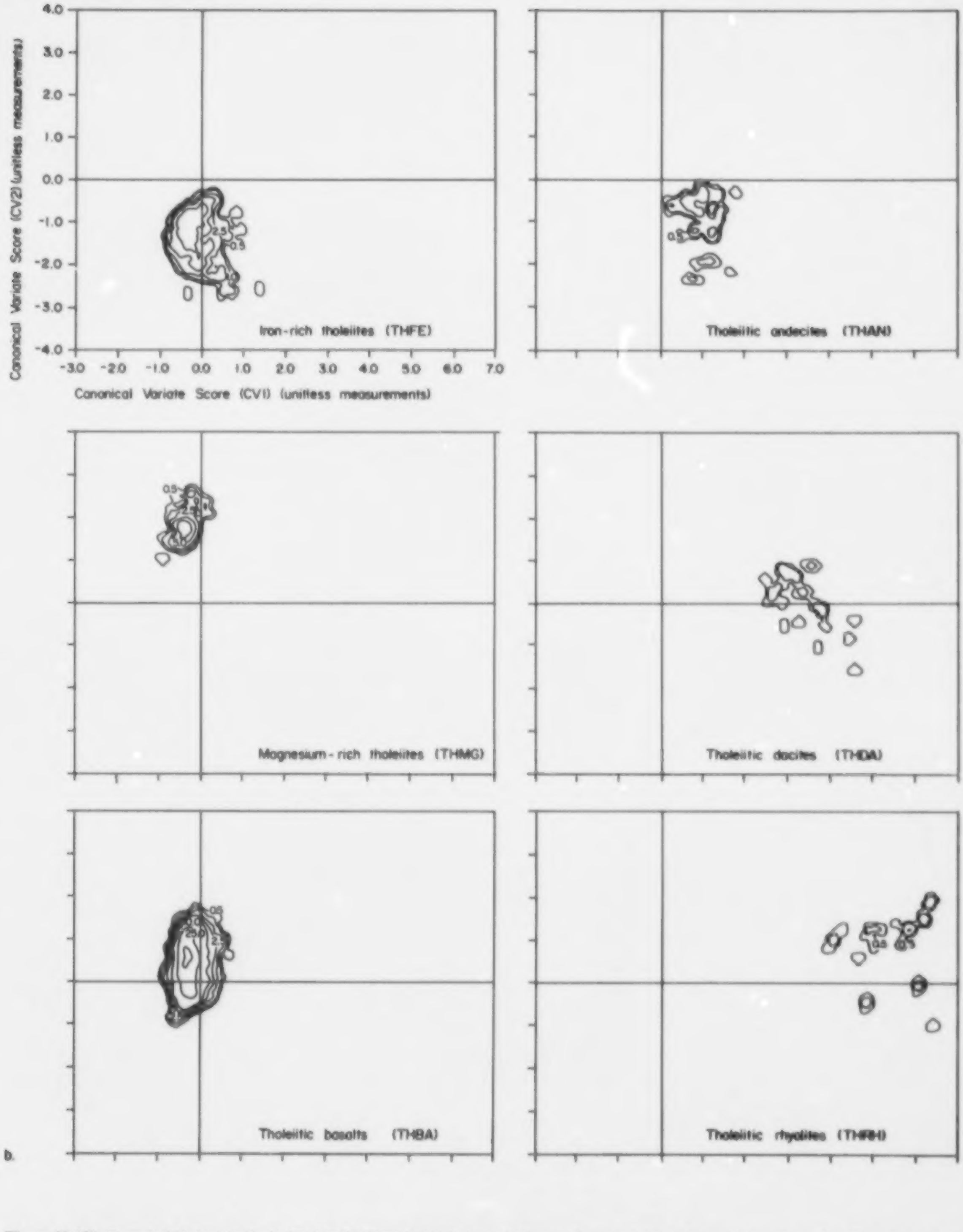

**Figure 28.18b.** Canonical Variate Analysis (CV1-CV2) PETROCH Tholeiitic Reference Groups, Additive Logratio transformation. Plot of canonical scores of the reference group samples (1112 samples). The magnesium-rich tholeiites (THMG) plot along the positive side of the CV2 axis and the iron-rich tholeiites (THFE) plot along the negative CV2 axis. Tholeiitic basalts (THBA) plot near the origin. The fractionated tholeiitic rocks (THAN, THDA, THRH) plot along the positive CV1 axis. Note that the fractionated rocks are more dispersed. Each reference group is contoured separately so that relative densities and overlap can be observed. The figure was divided into 100 cells along the x-axis, 80 cells along the y-axis. Contour intervals are: 0.5, 1, 2.5, 5, 10, 25 and 50 points per square unit.

**Table 28.12.** Canonical variate analysis—Calc-alkalic reference groups (CALC): Calc-alkalic basalt (CABA), calc-alkalic andesite (CAAN), calc-alkalic dacite (CADA) and calc-alkalic rhyolite (CARH).

**Chi-Square Tests With Successive Roots Removed**

| Roots removed | Eigenvalue | $\chi^2$ | % Trace |
|---|---|---|---|
| 0 | 11.052 | 1496 | 98.74 |
| 1 | 0.116 | 77 | 1.04 |
| 2 | 0.025 | 14 | 0.22 |

|  | Discriminant Function Coefficients |  |  | Canonical Scores |  |  | Communalities for 3 Discriminant Functions |
|---|---|---|---|---|---|---|---|
|  | DF1 | DF2 | DF3 | CV1 | CV2 | CV3 |  |
| Si | -0.4822 | 2.5421 | 2.0379 | -0.394 | -0.422 | -0.376 | 0.4748 |
| Al | -1.1757 | -0.1787 | -4.1518 | -0.247 | -0.465 | -0.474 | 0.5019 |
| Mg | 0.7469 | 0.5804 | 0.0604 | 0.615 | -0.376 | -0.340 | 0.6353 |
| Ca | 0.1255 | -0.9863 | -1.9722 | 0.190 | -0.400 | -0.602 | 0.5586 |
| Na | -0.1468 | -1.7459 | 1.3193 | -0.426 | -0.471 | -0.313 | 0.5009 |
| K | -0.0054 | -0.3116 | 0.4167 | -0.535 | -0.420 | -0.114 | 0.4761 |
| Ti | 0.0976 | 1.3933 | -1.6621 | 0.255 | -0.333 | -0.568 | 0.4980 |
| P | 0.0319 | -1.0046 | -0.0577 | 0.056 | -0.770 | -0.295 | 0.6828 |
| Mn | 0.1022 | 1.6794 | 0.2826 | 0.339 | -0.245 | -0.394 | 0.3302 |
| $Fe^T$ | 0.6643 | -2.6674 | 3.0694 | 0.421 | –0.443 | -0.369 | 0.5096 |

**Canonical Coordinates of Group Centroids**

|  | CV1 | CV2 | CV3 |
|---|---|---|---|
| Calc-alkalic basalt (CABA) | 0.7818 | 0.1997 | 0.0564 |
| Calc-alkalic andesite (CAAN) | 0.1473 | -0.3025 | -0.2364 |
| Calc-alkalic dacite (CADA) | -0.8369 | -0.4009 | 0.2101 |
| Calc-alkalic rhyolite (CARH) | -2.1440 | 0.5503 | -0.0977 |

Major element oxides are denoted by their elements only.
Wilk's $\Lambda = 0.0725$
F-ratio for $h^2$ (overall discrimination) = 42.81
N.d.f.1 = 56 and N.d.f.2 = 1659

total analysis including $Fe_2O_3$ and LOI. The analyses that were used, however, were the additive logratios with LOI as the divisor and $FeO^T$ used for iron.

Table 28.13b lists the Mahalanobis distances and posterior probabilities that have been calculated for each of these samples. Table 28.13b illustrates that the komatiitic rocks have a probability of 1.0 of belonging to the komatiitic reference group. For the same samples, there is a zero probability of belonging to either the tholeiitic or calc-alkalic reference group. Similarly, for the other samples, the probability of membership is highest in the reference clan from which these samples were taken. Exceptions to this are the fractionated tholeiitic rocks, which show some probability of group membership within the calc-alkalic clan. Sample 87MGM–1173 is a calc-alkalic basalt that also shows a posterior probability of being a tholeiitic rock. This illustrates the overlap between the calc-alkalic and tholeiitic basaltic rocks. In such a case, the decision to allocate to either one of the magma clans should be based on the sequence of volcanic rocks with which the sample is associated. Similarly, sample 82ECG–0260 has been classified as a tholeiitic rhyolite, but the posterior probabilities suggest that this could also be a calc-alkalic rhyolite. Again, the associated clan of rocks from where the sample was taken will assist in assigning the sample to the more correct rock type.

Table 28.13c shows the indices of typicality for the same representative samples. The probabilities are similar to Table 28.13b, however, in the case of the tholeiitic and calc-alkalic rocks, it is important to note that they have probabilities that are greater than zero for both reference groups. This indicates that there is some overlap between the group dispersions. The 2 samples, described above, also show indices of typicality that assigns them to more than 1 group. If field relationships are unavailable or uncertain, then the decision would be to allocate the sample to the group having the highest probability. If the probabilities between more than one group are nearly equal, then it is arbitrary as to which group the sample is allocated. This is most commonly observed between calc-alkalic and tholeiitic basaltic rocks where there is overlap between the dispersions as shown in Figure 28.16b.

Individual rock type reference groups can also be used to allocate unknown samples. Samples can be compared with the komatiitic suite (KOUM and KOBA), the tholeiitic suite (THFE, THMG, THBA, THAN, THDA and THRH) or the calc-alkalic suite (CABA, CAAN, CADA and CARH). Tables 28.14a and 28.14b show an example of allocation for 4 selected samples from the calc-alkalic reference group shown in Table 28.13. Table 28.14a lists the posterior probabilities; Table 28.14b lists the indices of typicality. Table 28.14a shows that the most of the samples have non-zero probabilities of membership in the reference groups from which they originate as well as in adjacent reference groups. This reflects the overlapping nature of the calc-alkalic groups as illustrated in Figure 28.14a.

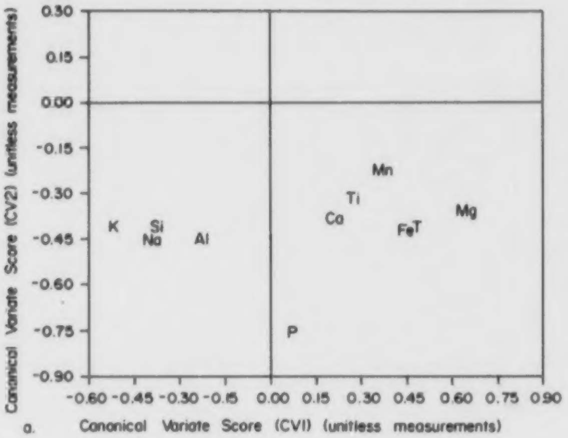

**Figure 28.19.** Canonical Variate Analysis (CV1-CV2) PETROCH Calc-Alkalic Reference Groups, Additive Logratio transformation. **a)** Plot of canonical scores of the major element oxides. Oxides are plotted as elements only. The relative positions of the elements indicate their relative relationships with respect to the magma clan reference groups. **b)** Plot of canonical scores of the reference group samples (578 samples). The calc-alkalic basalts and andesites (CABA, CAAN) plot along the positive side of the CV1 axis and the fractionated dacites and rhyolites (CADA, CARH) plot along the negative CV1 axis. Note that the fractionated rocks are more dispersed. Each reference group is contoured separately so that relative densities and overlap can be observed. The figure was divided into 100 cells along the x-axis, 80 cells along the y-axis. Contour intervals are: 0.5, 1, 2.5, 5, 10, 25 and 50 points per square unit.

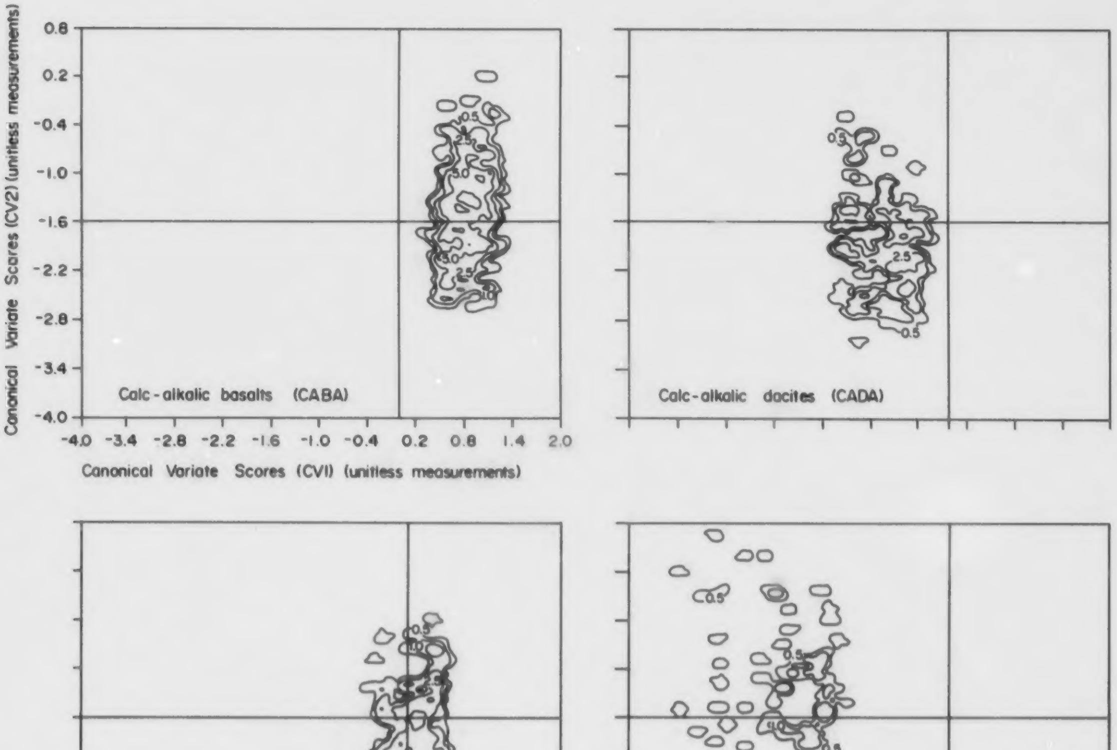

**Table 28.13a.** Compositions for selected samples from the 3 magma clan reference groups (komatiitic [KOM], tholeiitic [THOL] and calc-alkalic [CALC]).

| Sample | Reference Group | Rock Name | $SiO_2$ | $Al_2O_3$ | $Fe_2O_3$ | FeO | MgO | CaO | $Na_2O$ | $K_2O$ | $TiO_2$ | $P_2O_5$ | MnO | LOI |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 74J1759 | KOUM | Ultramafic | 45.90 | 6.57 | 0.00 | 10.80 | 20.90 | 9.04 | 0.44 | 0.07 | 0.44 | 0.03 | 0.24 | 4.20 |
| 79J02 | KOBA | Basaltic | 48.20 | 9.81 | 1.50 | 9.32 | 15.80 | 8.93 | 0.50 | 0.24 | 0.55 | 0.01 | 0.21 | 3.70 |
| 82ECG–1006 | THFE | Basalt | 48.70 | 12.90 | 5.47 | 9.13 | 6.12 | 9.74 | 3.15 | 0.32 | 1.44 | 0.06 | 0.17 | 0.60 |
| 84FWB–0034 | THMG | Mafic | 48.40 | 14.30 | 2.69 | 8.56 | 8.90 | 11.00 | 1.67 | 0.14 | 0.74 | 0.06 | 0.16 | 2.50 |
| 80RHS–0588 | THBA | Basalt | 49.50 | 14.50 | 0.00 | 11.88 | 7.14 | 9.87 | 2.38 | 0.14 | 1.02 | 0.07 | 0.27 | 1.50 |
| 82ECG–0164 | THAN | Basalt | 51.50 | 15.60 | 5.61 | 5.19 | 3.38 | 9.73 | 2.32 | 0.20 | 1.18 | 0.04 | 0.24 | 3.10 |
| 85NFT–7080 | THDA | Basalt | 62.40 | 13.50 | 1.01 | 7.11 | 2.17 | 5.21 | 3.49 | 1.25 | 0.66 | 0.09 | 0.25 | 0.90 |
| 82ECG–0260 | THRH | Rhyolite | 63.70 | 16.00 | 3.02 | 3.43 | 1.49 | 4.15 | 4.18 | 1.35 | 0.68 | 0.04 | 0.11 | 1.10 |
| 87MGM–1173 | CABA | Pyroclastic | 52.80 | 15.80 | 0.00 | 9.00 | 3.63 | 6.28 | 3.20 | 1.36 | 0.72 | 0.16 | 0.18 | 5.30 |
| 81CEB–1605 | CAAN | Andesite | 59.00 | 16.05 | 0.00 | 5.94 | 3.51 | 6.18 | 4.26 | 1.31 | 0.54 | 0.14 | 0.09 | 1.90 |
| 81GWJ–2049 | CADA | Int. Tuff | 63.40 | 15.90 | 1.76 | 2.34 | 2.27 | 5.78 | 4.10 | 1.33 | 0.49 | 0.17 | 0.07 | 1.10 |
| 83GWJ–4176 | CARH | Fragmental | 67.10 | 16.00 | 0.64 | 1.85 | 0.83 | 3.09 | 4.60 | 2.08 | 0.36 | 0.10 | 0.05 | 2.20 |

**Table 28.13b.** Posterior probabilities of selected samples from the 3 magma clan reference groups (komatiitic [KOM], tholeiitic [THOL] and calc-alkalic [CALC]).

| Sample | Reference Group | Komatiitic |  | Tholeiitic |  | Calc-Alkalic |  |
|---|---|---|---|---|---|---|---|
|  |  | D | PP | D | PP | D | PP |
| 74J1759 | KOUM | 4.6407 | 1.0000 | 46.3100 | 0.0000 | 57.6621 | 0.0000 |
| 79J02 | KOBA | 12.8561 | 1.0000 | 33.2482 | 0.0000 | 42.3109 | 0.0000 |
| 82ECG–1006 | THFE | 48.7317 | 0.0000 | 9.9302 | 0.9992 | 24.3155 | 0.0008 |
| 84FWB–0034 | THMG | 27.8060 | 0.0000 | 4.4236 | 0.9832 | 12.6115 | 0.0167 |
| 80RHS–0588 | THBA | 35.6178 | 0.0000 | 3.2054 | 0.9980 | 15.6912 | 0.0020 |
| 82ECG–0164 | THAN | 52.7017 | 0.0000 | 7.6846 | 0.9990 | 21.5367 | 0.0010 |
| 85NFT–7080 | THDA | 53.2636 | 0.0000 | 21.4271 | 0.9815 | 29.4951 | 0.0185 |
| 82ECG–0260 | THRH | 59.1928 | 0.0000 | 13.7217 | 0.6687 | 15.1422 | 0.3313 |
| 87MGM–1173 | CABA | 53.3390 | 0.0000 | 15.8715 | 0.3669 | 14.7749 | 0.6331 |
| 81CEB–1605 | CAAN | 49.7313 | 0.0000 | 15.4688 | 0.0033 | 4.0007 | 0.9967 |
| 81GWJ–2049 | CADA | 58.8378 | 0.0000 | 23.5777 | 0.0001 | 5.5059 | 0.9999 |
| 83GWJ–4176 | CARH | 72.3777 | 0.0000 | 30.1054 | 0.0000 | 7.4503 | 1.0000 |

D – Mahalanobis Distance
PP – Posterior Probability

**Table 28.13c.** Indices of typicality of selected samples from the 3 magma clan reference groups (komatiitic [KOM], tholeiitic [THOL] and calc-alkalic [CALC]).

| Sample | Reference Group | Komatiitic |  | Tholeiitic |  | Calc–Alkalic |  |
|---|---|---|---|---|---|---|---|
|  |  | D | IT | D | IT | D | IT |
| 74J1759 | KOUM | 0.4040 | 0.9453 | 17.2734 | 0.0000 | 21.8860 | 0.0000 |
| 79J02 | KOBA | 0.9200 | 0.5135 | 8.6556 | 0.0000 | 16.4754 | 0.0000 |
| 82ECG–1006 | THFE | 5.2784 | 0.0000 | 1.3503 | 0.1979 | 2.9343 | 0.0012 |
| 84FWB–0034 | THMG | 1.3886 | 0.1793 | 0.5554 | 0.8509 | 2.9835 | 0.0010 |
| 80RHS–0588 | THBA | 2.4798 | 0.0060 | 0.4885 | 0.8984 | 1.7523 | 0.0644 |
| 82ECG–0164 | THAN | 12.1611 | 0.0000 | 1.2263 | 0.2687 | 2.0200 | 0.0280 |
| 85NFT–7080 | THDA | 35.2211 | 0.0000 | 2.8322 | 0.0017 | 2.1653 | 0.0174 |
| 82ECG–0260 | THRH | 48.6624 | 0.0000 | 2.2782 | 0.0119 | 1.3586 | 0.1938 |
| 87MGM–1173 | CABA | 20.0857 | 0.0000 | 1.6365 | 0.0906 | 1.0912 | 0.3651 |
| 81CEB–1605 | CAAN | 25.8820 | 0.0000 | 2.1667 | 0.0174 | 0.4766 | 0.9060 |
| 81GWJ–2049 | CADA | 47.9891 | 0.0000 | 3.9348 | 0.0000 | 0.5672 | 0.8417 |
| 83GWJ–4176 | CARH | 109.9922 | 0.0000 | 5.8388 | 0.0000 | 0.5857 | 0.8269 |

D – Mahalanobis Distance
IT – Index of Typicality

**Table 28.14a.** Posterior probabilities of selected samples from the calc-alkalic reference groups (basalt [CABA], andesite [CAAN], dacite [CADA] and rhyolite [CARH]).

| Sample | Reference Group | Basalt D | Basalt PP | Andesite D | Andesite PP | Dacite D | Dacite PP | Rhyolite D | Rhyolite PP |
|---|---|---|---|---|---|---|---|---|---|
| 87MGM–1173 | CABA | 10.5109 | 0.8979 | 14.9815 | 0.1020 | 38.4465 | 0.0000 | 108.6629 | 0.0000 |
| 81CEB–1605 | CAAN | 10.5239 | 0.0463 | 4.3740 | 0.9491 | 15.2910 | 0.0046 | 69.8181 | 0.0000 |
| 81GWJ–2049 | CADA | 27.2731 | 0.0000 | 10.0356 | 0.1274 | 6.1058 | 0.8726 | 38.4319 | 0.0000 |
| 83GWJ–4176 | CARH | 82.2393 | 0.0000 | 47.8030 | 0.0000 | 13.5714 | 0.0140 | 4.8340 | 0.9860 |

D – Mahalanobis Distance
PP – Posterior Probability

**Table 28.14b.** Indices of typicality of selected samples from the calc-alkalic reference groups (basalt [CABA], andesite [CAAN], dacite [CADA] and rhyolite [CARH]).

| Sample | Reference Group | Basalt D | Basalt IT | Andesite D | Andesite IT | Dacite D | Dacite IT | Rhyolite D | Rhyolite IT |
|---|---|---|---|---|---|---|---|---|---|
| 87MGM–1173 | CABA | 1.4609 | 0.1480 | 2.5733 | 0.0043 | 4.0606 | 0.0000 | 4.9589 | 0.0000 |
| 81CEB–1605 | CAAN | 1.6481 | 0.0876 | 0.4854 | 0.9004 | 1.5715 | 0.1091 | 2.6065 | 0.0038 |
| 81GWJ–2049 | CADA | 5.7511 | 0.0000 | 1.6412 | 0.0894 | 0.5814 | 0.8304 | 1.5458 | 0.1172 |
| 83GWJ–4176 | CARH | 20.8678 | 0.0000 | 8.5038 | 0.0000 | 1.3864 | 0.1803 | 0.4217 | 0.9368 |

D – Mahalanobis Distance
IT – Index of Typicality

Table 28.14b lists the indices of typicality, which also show that the membership probability of the samples overlap into neighbouring groups.

Allocation procedures were applied to the samples that were rejected using the criteria discussed in the section entitled Data Selection. The samples were allocated using the covariances of the 3 main magma suites. The posterior probabilities indicate that the forced membership of most of the samples placed them into the reference group in which they were initially classified. Exceptions to this represent samples that are very atypical and cannot be properly classified. The compositions of these samples are very atypical and represent rocks whose compositions have been modified by metasomatism or alteration.

Indices of typicality were also calculated for the same samples. Most of the samples have indices equal to zero for the 3 reference groups. However, some samples also have non-zero probabilities. This suggests that they may belong to one of the reference groups. These samples may have been rejected because of unacceptable normative mineral values or because one of the major element oxides exceeded some specific value. However, the dispersion of the reference groups is such that the $\chi^2$ distance from these rejected values is still within the distance that produces a non-zero probability. This implies that the selection and/or rejection procedures could be further refined, or that these samples should be included in the reference group.

Allocation procedures thus provide a means of testing an unknown sample for its likelihood of belonging to none, one, or more than one of the reference groups. Samples that have non-zero probabilities of membership can be considered to be least altered and thus can be used for classification purposes. Samples that have zero probability of membership in any of the groups can be considered to be atypical or altered. If they are considered to be altered, then these samples can be investigated further for the type of alteration that the chemistry represents. This has obvious applications for mineral exploration strategies.

## CONCLUDING REMARKS

The results presented here have demonstrated that the major magma clans of Archean volcanic rocks are composed of unique chemical compositions with systematic variation that makes them distinctive from each other. The implementation of procedures to reject altered samples is a means by which individual and groups of samples can be tested for alteration. The assumption, of course, is that the selection procedures are valid. Although the authors have rejected alkalic rocks, as classified by Irvine and Baragar (1971), as being altered volcanic rocks, recent evidence has suggested that this may not be the case. If so, then the samples must be re-allocated and the reference groups must be redefined. The likelihood of alkalic and calc-alkalic groups being part of a continuous magmatic trend requires further investigation.

The logratio transformation approach advocated by Aitchison (1986) has assisted in overcoming the statistical problems that result from the presence of constant sum data. These logratio transformations are an effective way of determining the statistical characteristics of compositional data.

The establishment of the reference groups has been carried out primarily based on the normative mineral compositions and Q-Q plot characteristics of the data. Once these groups have been established, the use of canonical variate analysis tests for the statistical separation between the groups and indicates whether or not further classification or allocation can be carried out. In the case of the reference groups selected for this study, the authors have shown that the 3 magma clan reference groups (komatiitic, tholeiitic and calc-alkalic) are statistically distinct and that unknown samples can be assigned a likelihood of membership through posterior probabilities or typicalities. Similarly, within each magma clan suite, unknown samples can be assigned probabilities with respect to each rock type (i.e., tholeiitic basalts, andesites, dacites and rhyolites). The refinement of the groups into ideal multivariate normal distributions is impossible to achieve. This has been reflected in the results of the canonical variate analysis, particularly for the fractionated tholeiitic rocks. Further studies on specific fractionated tholeiitic sequences in the Archean may provide a better data base for the these rocks.

Typicality indices of the reference groups often show that samples can display affinities for all 3 magma suites, but usually 1 suite dominates over the others. Allocation applied to the altered samples has shown that most cannot be assigned membership within any of the reference groups. This is most often the result of alteration and not a failure of the method to classify. If classification is a must, then the posterior probability will give the membership through forced allocation.

For those samples that are correctly classified, the use of the standard petrological variation diagrams, such as those in Figures 28.5, 28.7 and 28.8, may be used with some degree of certainty. Nonetheless, it must be emphasized that these diagrams are difficult to interpret because of the lack of information regarding the other compositional elements. Aitchison (1990) proposed alternative graphical methods for describing and characterizing compositional relationships. A multivariate analysis may be preferable over a standard variation diagram because the reduced number of dimensions can provide much more information about compositional variation than a conventional variation diagram. The use of procedures such as logcontrast principal components analysis emphasizes maximum variance between the variables and is a useful method for investigating the simultaneous variation of the major element oxides in 1 or 2 diagrams. The use of canonical variate analysis emphasizes the maximum differences between reference groups and provides a measure of the uniqueness of the reference group.

It must be emphasized that this study is not intended to replace detailed petrographic work or provide a shortcut in determining the appropriate classification of an unknown sample. Rather, it is intended to be used in conjunction with suites of samples collected within an area and to provide a basis of comparison with the least altered reference groups that have been established. Most importantly, this approach cannot replace detailed knowledge of petrographic or field relationships that are critical in establishing relationships between petrochemical trends and tectonic environments that exist within geological domains.

The authors have established reference groups with typical compositions that can be used as a guide for the characterization of Archean volcanic rocks as shown in Table 28.3. The results presented in this chapter should not be considered to be complete. Further work needs to be carried out by examining the reference groups in more detail and by examining the amount of dispersion that should characterize each reference group. The use of the Q-Q plots may have resulted in reference groups that may be too restricted in compositional variation. Further work needs to be carried out on how much trimming should be done using these plots. As well, the selection of samples from the fractionated tholeiitic and calc-alkalic volcanic rocks needs to be re-addressed with the reference group selection being based on known tholeiitic and calc-alkalic sequences.

The samples that were rejected in the selection of the reference groups contain information on the nature of alteration and metasomatism associated with Archean volcanic rocks. From a mineral exploration point of view, the rejected samples form an important suite of samples that provide information regarding alteration associated with many types of mineralized environments. The approach adopted for this study can also be useful in the analysis of lithogeochemistry in mineral exploration strategies. Such an approach has already been used successfully by Smith et al. (1984) and Garrett (1989a, 1989b). Additional research is required into the characterization of metasomatic assemblages that characterize regional metamorphic compositions and alteration assemblages that characterize various types of ore deposits.

The use of additional data sets from PETROS (Mutschler et al. 1976) and IGBA (Li and Chayes 1983) may also assist in developing and extending reference groups for post–Archean rocks. The use of these more recent volcanic series can be used to compare the characteristics of Archean volcanic rocks with modern-day volcanic rocks. Examining the characteristics of individual greenstone sequences, greenstone belts and subprovince domains through multivariate data analysis and statistical techniques may provide a better understanding of the chemical variation and the magmatic evolution of the Archean crust.

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# APPENDIX 1

## Replacement of Zeros in a Composition

Given a compositional data vector $x$ with $D$ major element oxides and $C$ major element oxides equal to zero, each $x_i$ that is equal to zero is replaced by

$$x_i = \delta * (C+1) * (D-C)/D^2 \quad i = 1, .., C$$

and the non-zero elements are subsequently subtracted by

$$x_j = x_j - \delta * C * (C+1)/D^2 \quad j = C, .., D$$

where δ represents the maximum possible roundoff error for weight % data. For this study, a value of δ=0.005 (weight %) was chosen.

Full details are given in Aitchison (1986, p.268).

# APPENDIX 2

## Canonical Variate Analysis

The method is based on the following procedure. Given a data matrix $X$, composed of $N$ samples and $m$ variables, three 1 m by 1 m matrices are computed that are composed of the following relationships:

**T matrix:** sum of squares and products of all samples and variables (total covariance of all the data), defined as:

$$t_{ij} = \sum_{k=1}^{N} (x_{ik} - \bar{x}_i)(x_{jk} - \bar{x}_j)$$

where

$\bar{x}_i$ is the total sample mean of the $i^{th}$ major element oxide

$\bar{x}_j$ is the total sample mean of the $j^{th}$ major element oxide

$x_{ik}$ is the value of the $i^{th}$ major element oxide of the $k^{th}$ observation in the total sample

$x_{jk}$ is the value of the $j^{th}$ major element oxide of the $k^{th}$ observation in the total sample

**A matrix:** sum of squares and products of samples among each reference group (covariance among the different reference groups), defined as:

$$a_{ij} = \sum_{k}^{g} (\bar{x}_{ik} - \bar{x}_i)(\bar{x}_{jk} - \bar{x}_j)$$

where $g$ is the number of reference groups

**W matrix:** sum of squares and products of samples within each reference group (covariance within each reference group), defined as:

$$w_{ij} = \sum_{k=1}^{g} \sum_{m=1}^{N_k} (x_{imk} - \bar{x}_{ik})(x_{jmk} - \bar{x}_{jk})$$

where

$N_k$ is the number of samples in each reference group

$x_{imk}$ is the $i^{th}$ major element oxide of the $m^{th}$ observation in the $k^{th}$ reference group

$x_{jmk}$ is the $j^{th}$ major element oxide of the $m^{th}$ observation in the $k^{th}$ reference group

$\bar{x}_{ik}$ is the mean value of the $i^{th}$ major element oxide in the $k^{th}$ reference group

$\bar{x}_{jk}$ is the mean value of the $j^{th}$ major element oxide in the $k^{th}$ reference group

Given 2 or more distinct reference groups, the relationships between the covariances of the groups is described by:

$$\mathbf{T} = \mathbf{A} + \mathbf{W}$$

The method of canonical variate analysis is based on maximizing the ratio of the variation among the groups (**A**) to the variation within the groups (**W**). The number of canonical roots that can be obtained is g–1 groups.

The test statistic, Wilk's $\lambda$, is a measure of similarity of the 3 reference groups. As $\lambda \rightarrow 0$, the greater the distinction between the groups. As $\lambda \rightarrow 1$, the greater the similarity between the groups. This statistic measures the ratio of variation within the reference groups (**W**) to the total variation over all of the groups (**T**). The lower the number, the less likely that the groups share common characteristics. The generalized correlation ratio $\eta^2$ is simply $1-\lambda$ defined by **A/T**. This test assumes equality of the reference groups' covariance matrices. Frequently, this is not the case and the Wilk's $\lambda$ test must be interpreted with caution. The $\chi^2$ values provide a measure of the significance of each discriminant function. Values of $\chi^2$ that are less than an accepted level of significance (99%) indicate that the function may not be significant. The $\lambda$ associated with each function indicates how different the reference groups are with respect to the discriminant function. The trace indicates how much discriminating information is obtained from the discriminant function.

The use of the F-test allows the comparison of the similarities of the 3 groups. By assuming a level of significance (99% in this study) and determining the number of degree of freedom, a critical value of F can be determined. If the F-test score is less than the critical value, then it is likely, within the level of significance, that the groups are similar. If the F-test score is greater than the critical value, then as the F-test score increases, the more distinct the groups. The discriminant function coefficients (DF) provide the coefficients for functions that separate the reference groups. The canonical variate scores (CV) for each function give the co-ordinates in the discriminant function space for each variable and provide the relative relationships of the variables.

Caution must be applied using the canonical variate analysis (CVA) procedure. Samples that do not share any characteristics with the reference populations will still be classified into one of the discriminant reference groups. LeMaitre (1976a, 1982) discussed this problem and ways of overcoming it.

# APPENDIX 3

## Allocation (Classification) Procedures

The mathematics of allocation are straight forward and can be found in Cooley and Lohnes (1971), LeMaitre (1976a) and Garrett (1989b). Given a covariance (dispersion) matrix $S$, $k$ groups, with samples composed of $i$ variables, the $D^2$, the generalized Mahalanobis distance, is calculated such that

$$D_k^2 = (x_i - x_{ik})' S_k^{-1} (x_i - x_{ik})$$

where $k$ is the $k^{th}$ group

## Index of Typicality

For the index of typicality, the unknown sample is provisionally allocated to the $j^{th}$ of the $g$ groups such that

$$D_j^2 + \ln |S_j| = \min \; [D_k^2 + \ln |S_k|]$$

where $k = 1, ..., j, ... g$ reference groups

If the covariances of the groups are equal, then the logarithmic term is dropped.

The probability of group membership of each sample is predicted for each reference group using the statistic:

$$F = \frac{(N-g-p+1)n_k}{p\ (N-g)\ (n_k+1)}\ D_k^2$$

where

$N$ = total number of samples over the groups being tested

$g$ = number of groups

$n_k$ = number of samples in group $k$

$p$ = number of variables

This statistic is distributed as $F$ with $p$ (numerator) and $(N-g-p+1)$ denominator degrees of freedom. This is a "predictive" probability rather than an "estimative" probability that would normally be computed using the $\chi^2$ distribution. Estimative procedures have been shown to underestimate group membership probabilities (Garrett 1989a), thus the predictive approach is a better choice for classification.

## Posterior Probability

A detailed description of estimates of posterior probabilities can be found in Cooley and Lohnes (1971) and Campbell (1984). The estimation of posterior probability of membership in the $k^{th}$ population is given by:

$$pr(k; \bar{X}_m) = p_k f(\bar{X}_m; P_k) / \sum_{j=1}^{g} p_j\ f(\bar{X}_m; P_j)$$

where

$pr(k; \bar{X}_m)$ is the posterior probability of membership in the $k^{th}$ group with group vector (centroid) $\bar{X}_m$

$p_k$ is the prior probability that the unknown sample belongs to the $k^{th}$ group

$f(\bar{X}_m; P_k)$ is the value of the probability density function for the group vector $\bar{X}_m$.

The posterior probability is the probability of belonging to the $k^{th}$ group divided by the sum of the probabilities belonging to all $g$ groups. An unknown sample is then assigned to the reference group with the highest posterior probability (smallest Mahalanobis distance).

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## CONVERSION FACTORS FOR MEASUREMENTS IN ONTARIO GEOLOGICAL SURVEY PUBLICATIONS

| **Conversion from SI to Imperial** | | | **Conversion from Imperial to SI** | | |
|---|---|---|---|---|---|
| *SI Unit* | *Multiplied by* | *Gives* | *Imperial Unit* | *Multiplied by* | *Gives* |
| | | | LENGTH | | |
| 1 mm | 0.039 37 | inches | 1 inch | **25.4** | mm |
| 1 cm | 0.393 70 | inches | 1 inch | **2.54** | cm |
| 1 m | 3.280 84 | feet | 1 foot | **0.304 8** | m |
| 1 m | 0.049 709 7 | chains | 1 chain | 20.116 8 | m |
| 1 km | 0.621 371 | miles (statute) | 1 mile (statute) | **1.609 344** | km |
| | | | AREA | | |
| 1 cm$^2$ | 0.155 0 | square inches | 1 square inch | **6.451 6** | cm$^2$ |
| 1 m$^2$ | 10.763 9 | square feet | 1 square foot | **0.092 903 04** | m$^2$ |
| 1 km$^2$ | 0.386 10 | square miles | 1 square mile | 2.589 988 | km$^2$ |
| 1 ha | 2.471 054 | acres | 1 acre | 0.404 685 6 | ha |
| | | | VOLUME | | |
| 1 cm$^3$ | 0.061 02 | cubic inches | 1 cubic inch | **16.387 064** | cm$^3$ |
| 1 m$^3$ | 35.314 7 | cubic feet | 1 cubic foot | 0.028 316 85 | m$^3$ |
| 1 m$^3$ | 1.308 0 | cubic yards | 1 cubic yard | 0.764 555 | m$^3$ |
| | | | CAPACITY | | |
| 1 L | 1.759 755 | pints | 1 pint | 0.568 261 | L |
| 1 L | 0.879 877 | quarts | 1 quart | 1.136 522 | L |
| 1 L | 0.219 969 | gallons | 1 gallon | **4.546 090** | L |
| | | | MASS | | |
| 1 g | 0.035 273 96 | ounces (avdp) | 1 ounce (avdp) | 28.349 523 | g |
| 1 g | 0.032 150 75 | ounces (troy) | 1 ounce (troy) | **31.103 476 8** | g |
| 1 kg | 2.204 62 | pounds (avdp) | 1 pound (avdp) | **0.453 592 37** | kg |
| 1 kg | 0.001 102 3 | tons (short) | 1 ton (short) | **907.184 74** | kg |
| 1 t | 1.102 311 | tons (short) | 1 ton (short) | **0.907 184 74** | t |
| 1 kg | 0.000 984 21 | tons (long) | 1 ton (long) | **1016.046 908 8** | kg |
| 1 t | 0.984 206 5 | tons (long) | 1 ton (long) | **1.016 046 908 8** | t |
| | | | CONCENTRATION | | |
| 1 g/t | 0.029 166 6 | ounce (troy)/ ton (short) | 1 ounce (troy)/ ton (short) | 34.285 714 2 | g/t |
| 1 g/t | 0.583 333 33 | pennyweights/ ton (short) | 1 pennyweight/ ton (short) | 1.714 285 7 | g/t |

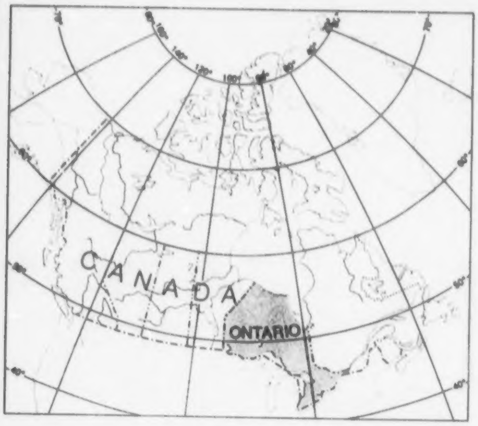
### OTHER USEFUL CONVERSION FACTORS

| | *Multiplied by* | |
|---|---|---|
| 1 ounce (troy) per ton (short) | 20.0 | pennyweights per ton (short) |
| 1 pennyweight per ton (short) | 0.05 | ounces (troy) per ton (short) |

*Note: Conversion factors which are in bold type are exact. The conversion factors have been taken from or have been derived from factors given in the Metric Practice Guide for the Canadian Mining and Metallurgical Industries, published by the Mining Association of Canada in co-operation with the Coal Association of Canada.*

# Map Themes

In celebration of the centennial of the provincial Mines and Minerals Division, the OGS has published a series of landmark geoscience maps of Ontario to accompany the Special Volume which consists of 2 parts. These map themes consist of 4 sheets—Northern, West-Central, East-Central, Southern—at a scale of 1:1 000 000.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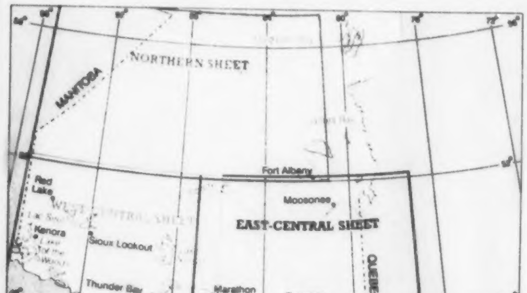
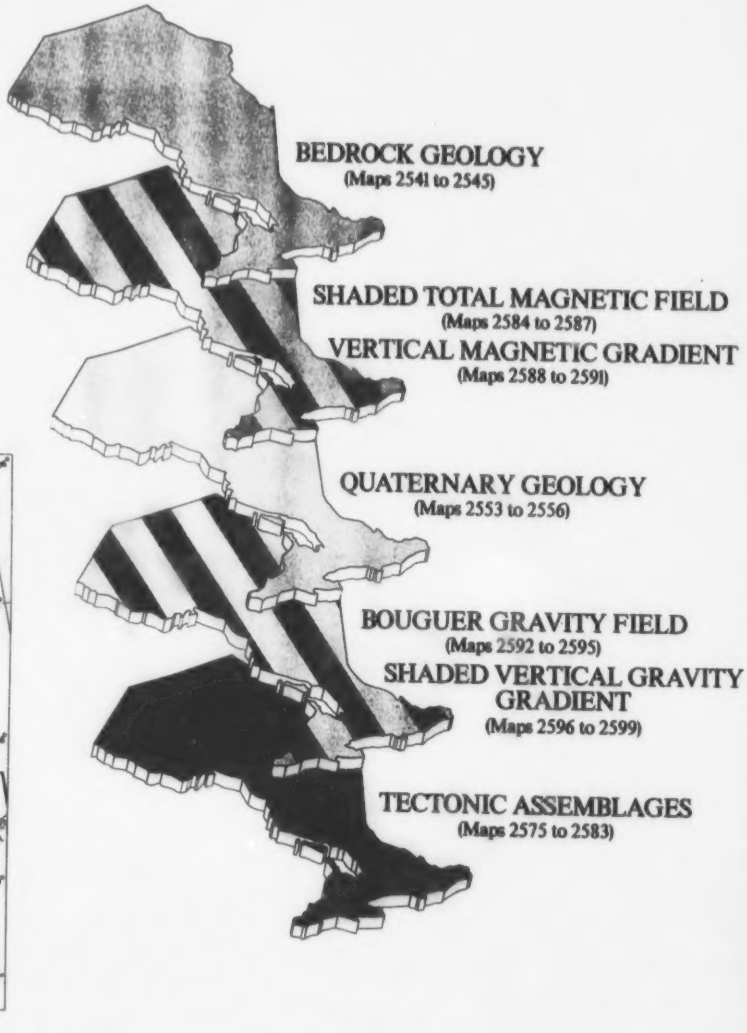

NORTHERN SHEET
MANITOBA
Fort Albany
Red Lake
WEST-CENTRAL SHEET
Kenora
Sioux Lookout
Moosonee
EAST-CENTRAL SHEET
QUEBEC
Thunder Bay
Marathon
Timmins
Kirkland Lake
U.S.A.
Lake Superior
Wawa
Cobalt
Sault Ste. Marie
Sudbury
Elliot Lake
SOUTHERN SHEET
Ottawa
Lake Michigan
Lake Huron
Lake Simcoe
St. Lawrence R.
Toronto
Lake Ontario
London
Windsor
Lake Erie
U.S.A.

Index to *Geology of Ontario* maps.


BEDROCK GEOLOGY
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